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ANCIENT ARCTICA[1]

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ABSTRACT

"Ancient Arctica" is the name here applied to the lands and seas of the past in the Arctic region. A survey was made of the shields that face and partly surround the Arctic Sea, of the intra-shield orogenic belts that extend into and under the sea, of the troughs of deposition and orogenic belts of Alaska, and of the topography of the Arctic Sea floor. It is concluded that the region is underlain by continental crustal material chiefly of shield character; that appreciable changes have occurred in the distribution of land and water by epeirogenic and orogenic movements of the crust; that the present deep basin of the Arctic Sea began to sink in Carboniferous time; that the Alaska–northeastern Siberia region was one of nearly constant land connection during Mesozoic and Cenozoic time and probably also during the Paleozoic; and that the crust of the North Atlantic and Greenland Sea underwent movements in Tertiary time sufficient to provide land bridges for migration of animals and plants between Europe and North America, possibly on several occasions.

Brief reference is made to the problems of biogeography, and the bearing that the theory of Ancient Arctica has on some of these problems.

The lands of the Arctic hold good petroleum possibilities, and those of the Arctic archipelago are singled out for special discussion because of their particular promise.

INTRODUCTION

DEFINITION OF ARCTICA

The word "Arctica" is here defined to include (1) Alaska, (2) the Arctic archipelago north of the North American continent, (3) the great island of Greenland, (4) the shelf areas adjacent to the continents that border the Arctic Sea, (5) the deep basin of the Arctic Sea, (6) the shield areas of the various continents that face the Arctic Sea, and (7) the orogenic belts that impinge on the Arctic Sea or extend into and under it (fig. 1). A review of the geology of the lands of Arctica and of the topography of the sea floor points to the existence from time to time of land in places where the waters now spread. To the lands and seas of the north polar region of the geologic past the name "Ancient Arctica" is given.

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PURPOSE OF PAPER

The purpose of this paper is to review the geology of Arctica in order to pursue the problem of the relation of Eurasia to North America and Greenland. The conclusions will have an obvious bearing on the problems of biogeography, both past and present. They may also have an important bearing on exploration for oil and gas in the Arctic.

ALASKA

PALEOZOIC GEOSYNCLINE AND RELATED OROGENY

Most of the Paleozoic rocks of Alaska are exposed in the Brooks Range, Seward Peninsula, the central Yukon drainage

area (Central Plateau), and the Alexander Archipelago (fig. 2). The Alaska, Nutzotin, and Wrangell ranges also contain Paleozoic rocks, and, near by, a belt extends along part of Copper River and Chitina River valleys. The towering mass of Mount McKinley in the Alaska Range is eroded mostly from deformed Paleozoic strata.

For a detailed study of the Paleozoic rocks of Alaska, Smith's U.S. Geological Survey Professional Paper 192 should be consulted, particularly the large correlation chart in the pocket. The formations are well exposed in the Central Plateau, and a résumé of a few selected sections is given in table 1.

The igneous rocks in the Yukon-Tanana region have been summarized by Mertie (1935, p. 302) as follows: Basic lavas of basaltic and diabasic character were extruded during at least five geologic epochs in the Paleozoic. The first was in the Middle Ordovician, the second in the Middle Devonian, and the last three during three epochs of the Carboniferous. Granular intrusives of the same general character accompanied the extrusion of the lavas, but the volume of such rocks is relatively small. Some rhyolite and dacite lavas and tuffs are found among the Carboniferous lavas, but, in general, lavas of acidic or intermediate character are rare. Ultrabasic rocks were intruded during the Upper Devonian epoch.

The volcanism that, according to Mertie, occurred during the Carboniferous period in Alaska was greater than in any other period and most intense in the Alaska Range. Eruptions of basic lavas accompanied epeirogenic movements that persisted into the Triassic.

It is immediately clear that the rocks of part of the Central Plateau and of the southern part of Alaska represent the volcanic archipelago assemblage previously recognized and described in the western Cordillera of southeastern Alaska, British Columbia, Washington, Oregon, California, and Nevada (Kay, 1947, p. 1290; and Eardley, 1947, p. 342). The presence of the basic intrusives in the volcanic assemblage suggests that the belt was the site of both intrusive and extrusive activity and not simply a trough adjacent to a volcanic archipelago.

In northern Alaska the Paleozoic rocks are mainly sandstones, shales, and limestones and are typical of the mainland assemblage, also previously described in the western Cordillera. No volcanic rocks have been found in the sediments.

If a line is drawn separating the volcanic archipelago assemblage from the mainland assemblage, it will appear somewhat like that of figure 1. The curvature of the present mountain systems has influenced the curvature given by the writer to the line, but, then, there is considerable value to the premise that the belts of deformation trend approximately parallel with the former troughs of sedimentation.

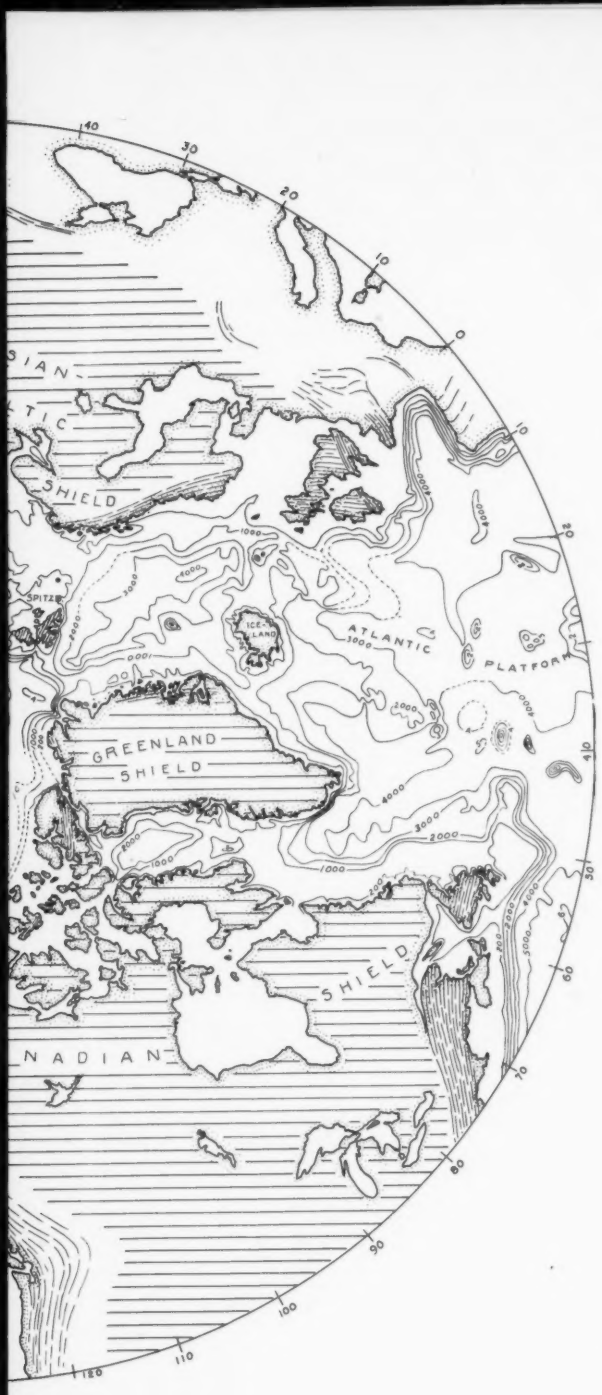
The distribution in outcrop and the geosynclinal thicknesses, where known, of the Paleozoic strata indicate that the whole of Alaska was a region of subsidence and sedimentation in the Paleozoic as an extension of the Cordilleran geosyncline. Where did the sediments of the mainland assemblage come from? The only possible source must have lain to the north where the Arctic Sea now spreads.

TRIASSIC AND JURASSIC GEANTICLINE AND ADJACENT BASINS

During the Triassic period and persisting into the Jurassic a great geanticline rose from the Paleozoic geosyncline and separated two adjacent basins of ac-

140
130

FIG. 1.—Paleozoic shields and orogenic belts adjacent to the Arctic Sea. Tectonics of Asia taken fro
contoured in meters from all available sources. Belts of parallel dashes are the orogenic belts. The shields a
Tschuktschen" is supposed to be a Paleozoic shield (Umbgrove).



Tectonics of Asia taken from Umbgrove (1947). North Polar seas orogenic belts. The shields are horizontally ruled. The "Nucleus of

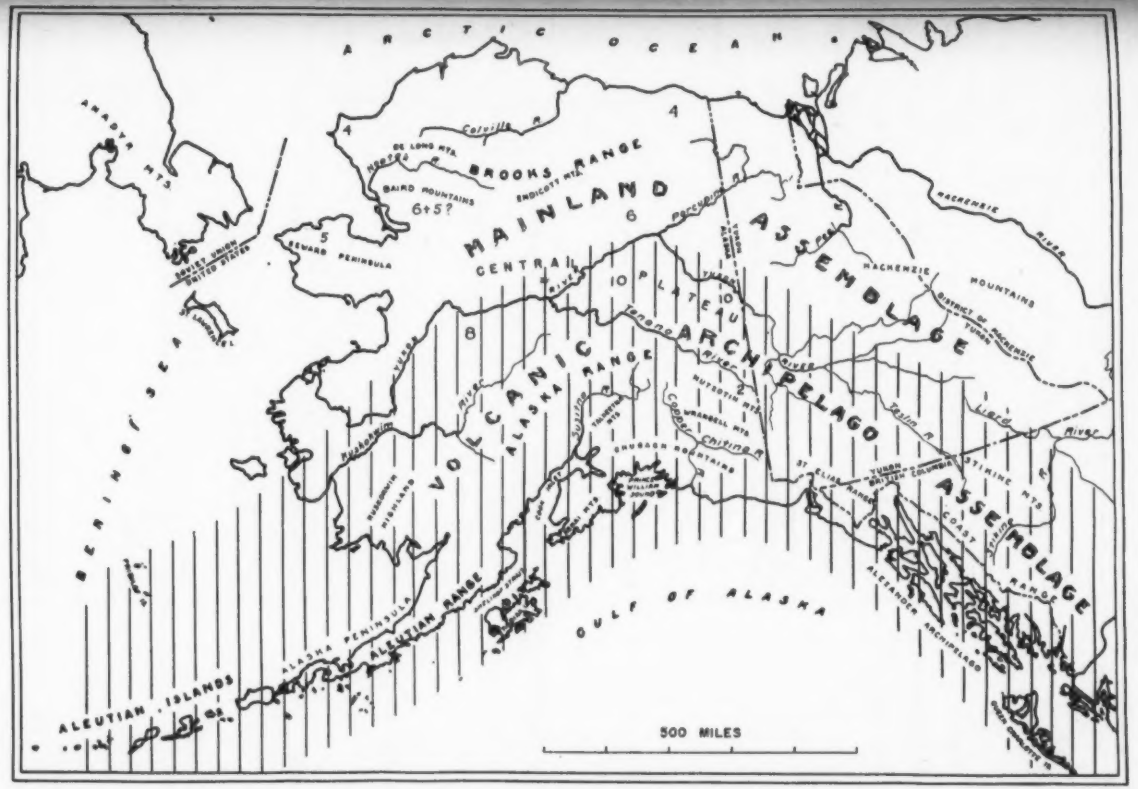


FIG. 2.—Alaskan Paleozoic geosynclinal divisions. Lined area is the volcanic archipelago assemblage, and the blank area is the mainland assemblage. Numbers represent thousands of feet of sediments and volcanics in those few places where estimates have been made. Generally, they represent only part of the Paleozoic section.

TABLE 1

SELECTED PALEOZOIC SECTIONS OF ALASKA*

Section	Description	Thickness
	Permian	
Kandik district	Limestone, conglomerate, shale and sandstone; Tahkandit limestone	.....
	Mississippian	
Porcupine district	Dark shale and limestone, in part same as Calico Bluff formation	.....
Koyukuk-Melozi district	Greenstone, little rhyolite, formerly considered part of Kanuti group	.....
Yukon-Tanana district	Clay shale, sandstone, conglomerate; Nation River formation	4,000– 6,000 feet
	Lava flows and associated sediments; Rampart group and Circle volcanics	5,000–10,000 feet
	Limestone, shale, slate; Calico Bluff formation Limestone beds Undifferentiated schist, shale, chert, quartzite	13,000 feet
	Chert with minor amounts of limestone and shale; Livengood chert	2,000– 4,000 feet
	Upper Devonian	
Eagle district	Basalt lava and pyroclastics of greenstone habit; Woodchopper volcanics	10,000± feet
	Middle Devonian	
Kantishna Nenana district	Massive limestone, equivalent to part of Tonzona	.....
	Limy shale, more calcareous at top	.....
	Shale, argillite, graywacke, quartzite	10,000± feet
	Black conglomerate, white conglomerate, shale, and graywacke	.....
	Middle Silurian	
Porcupine district	Black fissile shale, little siliceous limestone Buff magnesian limestone	2,500+ feet
	Middle Ordovician	
Ruby district	Magnesian limestone overlain by calcareous limestone	5,000 feet
Preacher district	Volcanic tuff and associated igneous rocks	.....
	Black shale, merging downward into schist	.....
	Upper Cambrian	
Kandik district	Limestone, with dark gray to black slate and chert in higher part	.....
	Limestone	.....
	Middle Cambrian	
Kandik district	Upper plate of limestone	.....
	Thin layers of slate and quartzite	.....
	Lower plate of limestone	.....

* Stratigraphic succession of the formations listed under any one period is not implied. The relative age is commonly not known.

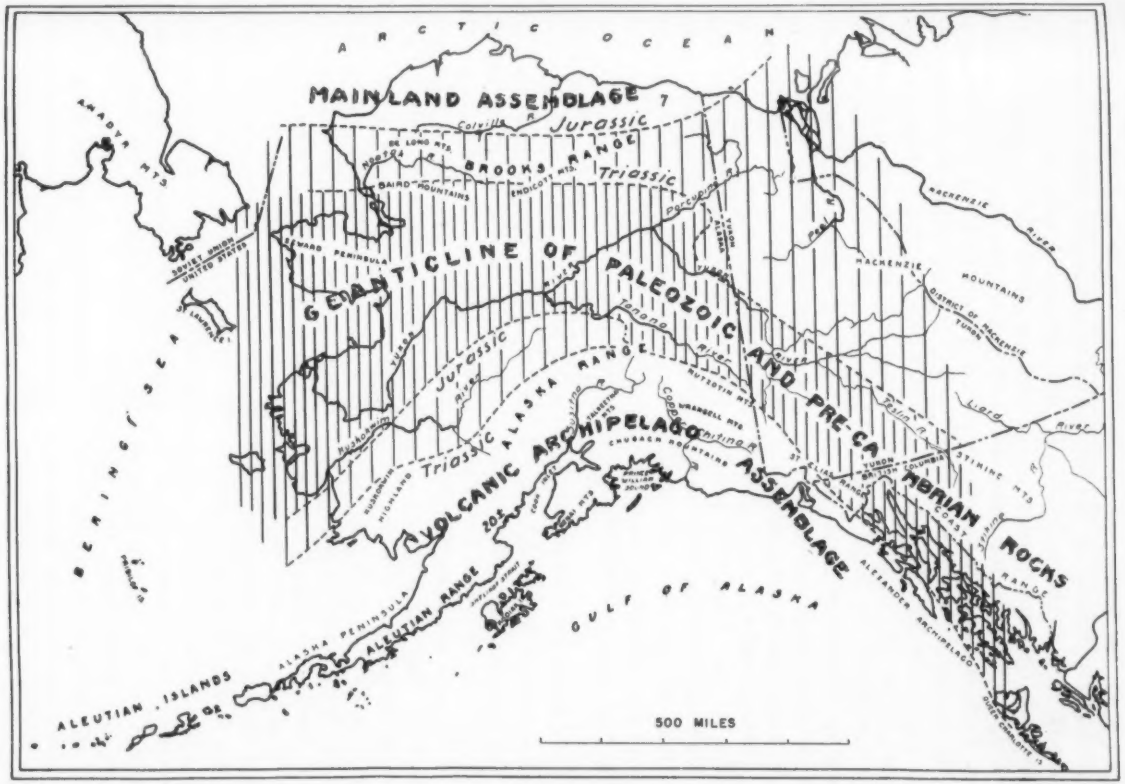


FIG. 3.—Alaskan geanticline of Triassic and Jurassic times and the regions of sedimentation on either side. Vertical ruling covers area of geanticline, which was somewhat farther north in Jurassic time than in Triassic. The close ruling is the area common to both Jurassic and Triassic uplifts, the open ruling on the north shows approximately the limit of the Jurassic uplift, and the open ruling on the south the limit of the Triassic uplift.

cumulation, one on the north and one on the south (fig. 3). The basin on the south collected chiefly sediments of the volcanic assemblage, while the trough on the north received limestones, sandstones, shales, and cherts of the mainland assemblage. Published maps do not exactly delimit the geanticline, but Mertie (1930, pp. 106–110) describes the region of epeirogenic uplift and erosion. The geanticline is analogous, except in detail, to the Mesozoic Cordilleran geanticline in Canada and the United States.

The Mesozoic geanticline that rose in the approximate center of the Paleozoic geosyncline of the Cordillera of the United States and Canada is so characteristic that its presence in Alaska strengthens the concept of an equally broad Paleozoic geosyncline there.

CRETACEOUS BASINS AND MOUNTAIN BELTS

An examination of the geologic map of Alaska and the correlation chart in Professional Paper 192 shows the Lower and Upper Cretaceous strata so widespread that much of Alaska must have been under water and receiving sediments during Cretaceous times. Certainly, large parts of the Triassic and Jurassic geanticline were covered. Mertie (1930, p. 112) states that at least at one time, perhaps at several, during the Cretaceous, all of Alaska was subjected to sedimentation. But he also concludes, because of the coarse clastic nature of most of the Cretaceous beds and the unconformities at their base and within them, that they were due to "differential warping" and even to mountain building. Linear uplifts and troughlike basins between seem to be discernible if the Cretaceous areas of the Territory are zoned off from the areas devoid of Cretaceous (fig. 4). The Brooks Range probably rose sharply at the end of Lower Cretaceous time and again after Upper Cretaceous time (Mertie, 1930, p. 124). The great Cordilleran geanticline of Canada enters Alaska and appears to split into four prongs, with the Brooks Range constituting the northern one, the Yukon River area from Fort Yukon to Ruby and beyond another, the Kantishna-Kuskokwim Bay belt a third, and the Nutzotin and Alaska ranges the fourth. South of this southern arcuate uplift a long narrow trough formed, in which a series of conglomerates, tuffs, black clays (now slates and phyllites), graywackes, limestones, argillites, and sandstones accumulated. Another uplift bordered this trough on the south which may have connected with the bordering archipelago of southeastern Alaska, British Columbia, and the United States.

Along the north side of the Alaska Range is a group of continental deposits of late Upper Cretaceous age, the Cantwell formation. It has a conspicuous basal conglomerate with boulders up to 6 inches in diameter; and there are many beds of conglomerate throughout the sequence. The formation is several thousand feet thick. It rests unconformably on other Cretaceous beds as well as on pre-Cretaceous rock. According to Mertie (1930, p. 116), the Cantwell was deposited during the first vigorous uplift of the Alaska Range, but the distribution of the pre-Cantwell Cretaceous beds indicates that the range had already been defined by earlier Cretaceous movements.

North of the Brooks Range the Lower Cretaceous has been estimated to be at least 10,000 feet thick, and the Upper Cretaceous at least 15,000 feet (Smith, 1939, p. 55). If they have these thicknesses where superposed, the trough was truly one of geosynclinal proportions. Perhaps enough land lay to the south to

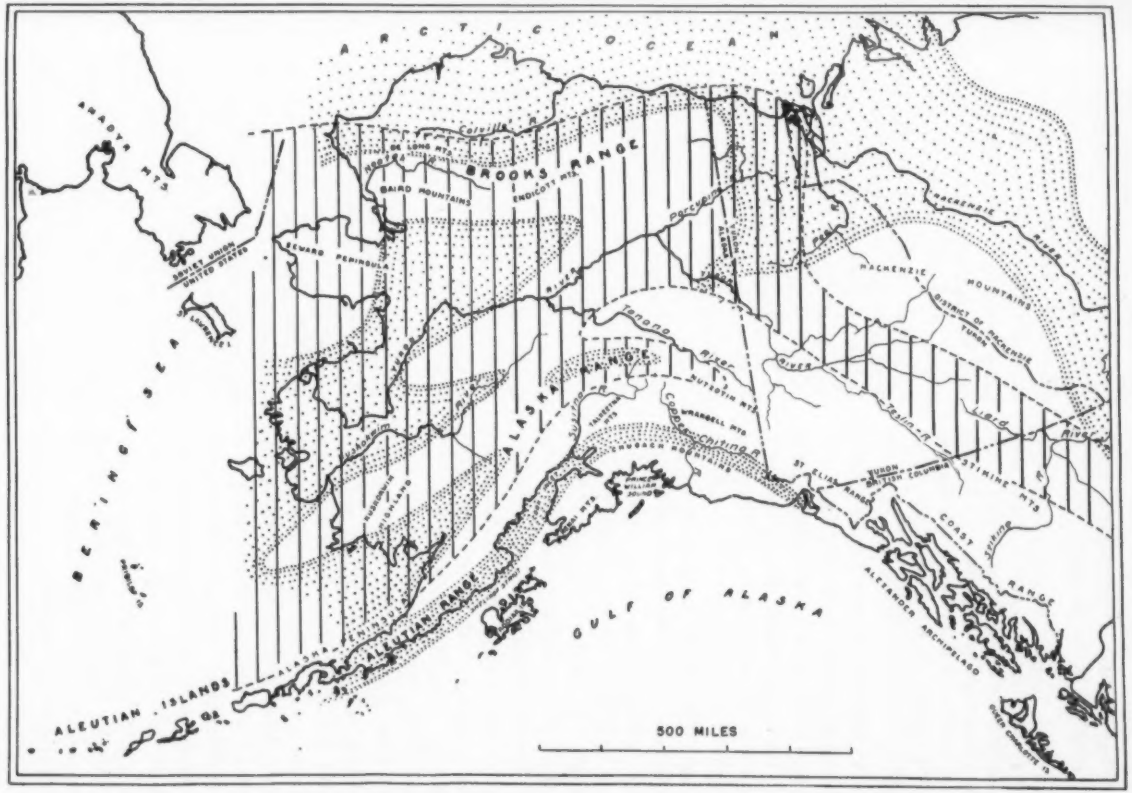


FIG. 4.—Cretaceous seas and linear uplifts and the later Laramide belt of compression. The seas are stippled, the uplifts are blank except where covered by the Laramide belt, and the Laramide belt is ruled. Boundaries are all approximate and based on available information.

furnish all the sediments; but, assuming the distribution shown in figure 4, it seems that at least some doubt is justified. Payne (1948, p. 36) says,

> . . . a thick Lower Cretaceous miogeosynclinal graywacke-shale sequence, without chert and volcanics, formed along the present northern foothills of the Brooks Range. . . . Deposition was rapid and marine in a foredeep between the source in a rising island arc to the south and stable craton to the north.

The Cretaceous in the Lower Yukon embayment is also thick and has been estimated to be over 10,000 feet. Except for the open end toward the Bering Sea, the basin is surrounded by uplifts from which the sediments must have come.

In both the northern Alaskan trough and the Yukon embayment an unconformity between Lower and Upper Cretaceous beds leads Mertie (1930, p. 124) to visualize appreciable crustal disturbance throughout the region at the close of the Lower Cretaceous. According to Payne (1948, p. 37), the Middle Cretaceous orogeny centered in the Baird Mountains of the Brooks Range.

The Cretaceous as a whole is lacking in volcanics except for tuffs; and the volcanoes that emitted the dust for the tuffs probably lay along the southernmost archipelago, which today is the site of the coast ranges or similar elements now under the sea on the continental shelf.

Intrusive igneous rocks are widespread over Alaska, but not enough is yet known to classify them as to age. One group, probably the largest, are Mesozoic in age. In the Alaska Range, in particular, most of them seem to be of late Cretaceous and early Tertiary age (Smith, 1939, p. 90).

LATE CRETACEOUS AND EARLY(?) TERTIARY OROGENIES

Other orogenic deposits within and north of the Alaska Range, probably younger than the Cantwell and of Eocene age, attest later phases of orogeny. The "coal formation" may overlie the Cantwell unconformably and is itself highly deformed. The Kenai coal-bearing formation south of the Alaska Range contains numerous volcanic rocks, such as breccias, agglomerates, tuffs and flows of basalt, andesite, trachyte, and rhyolite, and suggests crustal unrest; but for the most part no fossils have been found that enable accurate age assignment. It is possible that the volcanic archipelago that confined the Cretaceous trough along the Gulf of Alaska still existed and furnished most of the volcanics to the southern Eocene (?) zone of accumulation.

Above the coal measures north of the Alaska Range is the Nenana gravel, a coarse, well-rounded pebble and boulder deposit, some 2,000 feet thick. Locally it rests unconformably on the coal formation and may be the flanking waste of the newly built mountains. If the gravels are Miocene or Pliocene rather than Eocene in age, then there are two possibilities: first, the orogeny that deformed the coal measures is itself Eocene and was followed by a long period of erosion, after which epeirogenic uplift and the deposition of the gravels took place. On the other hand, the orogeny may be Miocene or Pliocene and may be responsible for deposition of the gravels. These possibilities have been outlined by Mertie (1930, p. 120), who points out also that the Cantwell orogeny was confined to southern Alaska and the Brooks Range but that the second phase (producing the Nenana gravels) spread over the Yukon as well. Both the Upper Cretaceous and the Eocene in the Yukon Basin have been deformed by open folding, and both have been intruded by monzonitic and basic rocks and covered in places by genetically related extrusives.

The site of the Aleutian Range was mostly submerged during the Cretaceous while the Alaskan Range was being outlined, but its late Cretaceous and early Tertiary history is similar to that of the Alaska Range. In later Tertiary times it became involved in Coast Range orogeny.

PLIOCENE AND PLEISTOCENE OROGENY

Areas of deposition and belt of disturbance.—The southern or Coast Ranges were the sites of the most modern orogeny in Alaska. They consist of Kodiak Island, the Kenai Peninsula, the Chugach Mountains, and the coastal part or more of the St. Elias Mountains. They form a partially submerged, arcuate mountain belt that faces the Gulf of Alaska. The belt consists of Mesozoic strata, largely undifferentiated, and of numerous Laramide (?) intrusions; but numerous Tertiary deposits record its history in late geologic times. Stretching from Lituya Bay to Katalla are outcrops of middle to late Tertiary strata, some sections very thick, and all tilted and folded (Smith, 1939, p. 59).

Quoting from Smith (1939, p. 60):

> As an indication of the former greater extent of these Tertiary rocks the isolated occurrence of them on Middleton Island, far out in the Pacific Ocean, is of particular significance. The rocks on this island are described by Capps as moderately indurated sandstone and conglomerate that in places are inclined 30° or more and have been beveled off by wave erosion, and later gravel and sand have been laid down on the truncated edges.

A thick Tertiary section in the Alaska Peninsula, from Kamishak Bay in the northeast to Pavlof Bay in the southwest, is mostly of volcanic origin. Some of the sediments are probably of Eocene age; others are probably Miocene. They are both terrigenous and marine and in places are probably 5,000 feet thick. Near Herenden Bay, Pliocene beds may occur. In the vicinity of Kodiak Island and on Trinity Island evidence of Miocene and Pliocene beds has been found. All the Tertiary rocks have been folded, dips of the recognized Miocene and Pliocene beds averaging 15°–20° and those of the older Tertiary beds usually much steeper.

By examination of the correlation chart and Geologic Map of Alaska in Professional Paper 192 it will be seen that the younger Tertiary rocks are all confined to the Coast Ranges except some in the Alaska Peninsula. The Coast Ranges were, therefore, the site of subsidence and sedimentation in Miocene and Pliocene time (fig. 5). Beds as young as Pleistocene probably accumulated at the eastern end. Because the strata are tilted and folded, it follows that the same belt became one of mountain building, perhaps during the deposition of the sediments as well as afterward. That the folding was gentler than in the Alaska Range during earlier orogenies is shown by the open character of most of the folds. If the Nenana gravel of the interior of Alaska is of late Tertiary age, it would appear that the arc of the Alaska Range was uplifted at the time of the Coast Range orogeny.

Because the Coast Range orogeny is so recent, the distribution of the mountains indicates the extent of the orogenic belt. The areas of sedimentation of Miocene and Pliocene time, shown by stippling on figure 5, are restricted to the present coastal margins and the continental shelf, whereas the belt of orogeny, shown by ruling, cuts farther inland to include the entire zone of Coast Ranges.

The map of figure 5 shows the relation of the late Tertiary coastal deposits of the Peninsula of Lower California, California, Oregon, Washington, British Co-

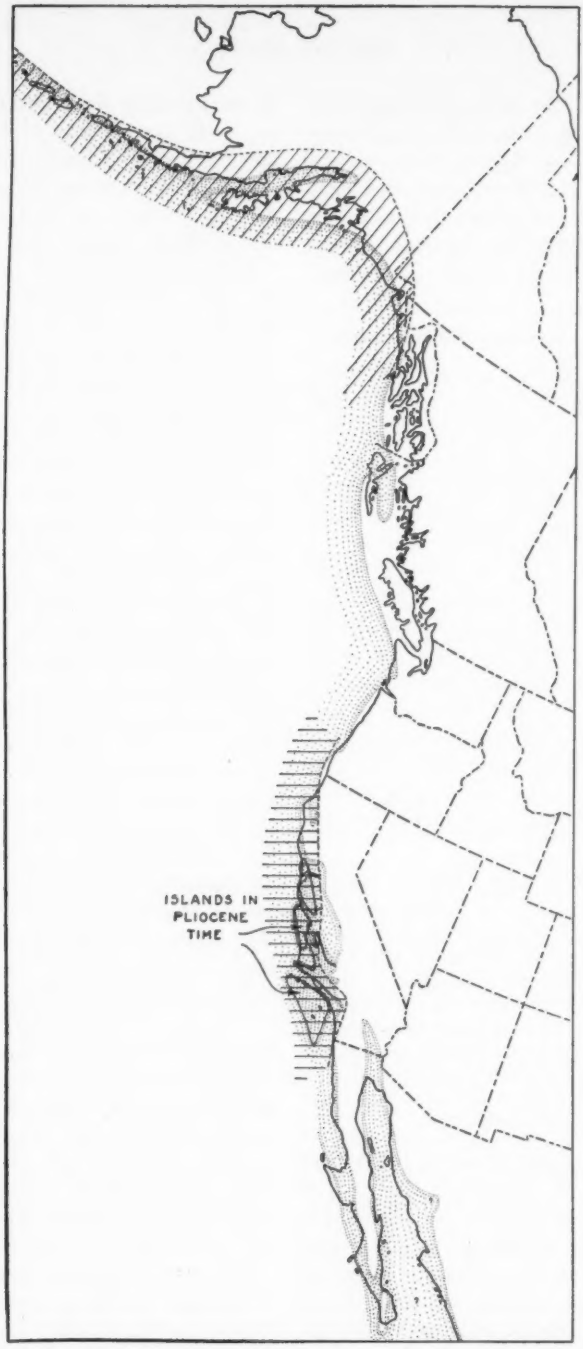


FIG. 5.—Comparison of Alaskan Miocene and Pliocene coastal geology with that of British Columbia, Washington, Oregon, California, and Mexico. Stippled areas represent Miocene and Pliocene deposits, and ruled areas belts of late Pliocene or Pleistocene compressional orogeny, where known. The Alaskan and Californian belts may be continuous but beneath water on the continental shelf.

lumbia, and southeastern Alaska to those of the Coast Ranges of Alaska. Intense compressional deformation in late Cenozoic time was localized in the central and northern Coast Ranges of California (map, fig. 5). The gentle compressive movements which started in the Miocene and then relaxed for a while surged to a peak in the late Pliocene and again to another peak in the mid-Pleistocene. The orogenic belt, if it continued northward, lies seaward of Oregon and Washington and, aside from gentle warping in the Island and Coast Ranges of British Columbia and southeastern Alaska, is not recognized again until the Gulf of Alaska coast is reached at Lituya Bay.

Modern volcanic belt.—A great arc of active or only recently dormant volcanoes extends from Mount Edgecumbe on Kruzof Island near Sitka through Mount Wrangell and the whole Alaska Peninsula and the Aleutian Islands to Kiska (fig. 6). This arc is 2,500 miles long, and many of the peaks rise 8,000–11,000 feet above the sea.

Southward from the Mount Spur group at the extreme northeastern limit of southwestern Alaska, the signs of Tertiary to Recent volcanism become increasingly evident until, south of Chignik, they make up practically all the features of the bedrock. The lofty modern volcanoes that overshadow all the other topographic features are dominant in almost every landscape. According to Smith (1939, p. 82):

> Their general features may be summarized by saying that the volcanic activity in the region as a whole seems to have begun somewhat after the beginning of the Tertiary and to have persisted intermittently to the present time. The composition of the lavas has in the main been fairly comparable with that of normal andesites, but more basic phases analogous to basalt and more acidic phases approaching rhyolite are by no means unknown. Practically every kind of volcanic activity, from stupendous explosion to quiet welling forth of lava, is represented in different areas and even in many individual volcanoes.

The arcuate belt of volcanoes is undoubtedly an active orogenic belt and closely related to the late Cenozoic belt of orogeny. Compare the map showing the volcanoes (fig. 6) with the one showing the Miocene-Pliocene deposits and the Coast Range orogenic belt (fig. 5).

CENOZOIC LAND CONNECTIONS WITH SIBERIA

The Cenozoic history of Alaska is similar in many respects to that of the great western Cordillera of Canada and the United States. It seems certain that, at least during Mesozoic and Cenozoic time, the Cordillera, with its troughs, geanticlines, compressional mountain belts, and volcanic arcs, extended into Siberia and there continued down the coast of Asia with branches inland. Therefore, since the Paleozoic, land and shallow seaways almost continuously connected North America with Asia, and the same conditions probably held during the entire Paleozoic era (Eardley, 1947).

As to the region north of the Brooks Range, now the Coastal Plain and Arctic Sea, it has been referred to by Payne as the "craton," by which he means the central stable region or shield of the continent. It is true that the Brooks Range, the foothill belt, and the Arctic Coastal Plain correspond to our frontal Rockies and Great Plains, but, on the other hand, it has not been proved that any of the sediments of the northern Alaska trough came from the north. They seem to have come from the Triassic and Jurassic geanticline and from the early Brooks Range of Middle Cretaceous age, both on the south. Since marine units are present in all the systems of the Arctic Coastal Plain, it would appear that the

FIG. 6.—Distribution of volcanoes in Alaska. Taken from Smith (1939). Dots are active or recently active volcanoes.

Arctic Sea made its appearance in early Mesozoic time and that its floor continued to subside thereafter. In the next section the geology of the Arctic Archipelago will be considered, and it is there concluded that large parts of the Arctic became seaways in Carboniferous time and that the Arctic Sea basin began to sink then. It does not appear, therefore, that an extensive land can be postulated directly north of Alaska after Carboniferous time. This subject will be considered at greater length later.

ARCTIC ARCHIPELAGO

INTRODUCTION AND LITERATURE

The Arctic Islands of Canada, together with Boothia and Melville peninsulas, form a geographic unit known as the "Arctic Archipelago." They extend from Hudson Bay at 62° north latitude northward for 1,500 miles to the northern tip of Ellesmere Island at 83° north latitude. Their greatest extent from east to west is about 1,000 miles.

Little is known of the hinterlands of the Arctic Islands because exploration has been confined, with but few exceptions, mainly to their coasts. However, sufficient information has been gathered to provide a general picture of the geography and geology, though no doubt this will be greatly modified by future more detailed explorations.

The geology of the Arctic Archipelago has been summarized recently by Armstrong (1947, pp. 311–324), and in the following pages the factual information has been taken chiefly from his writing. A report on Victoria Island and vicinity by Washburn (1947) was also helpful. Both reports stress the inadequate information on the vast region. The new Geologic Map of Canada should be consulted for place names used in the following paragraphs.

PRE-CAMBRIAN AND EARLY PALEOZOIC AREAS

Two large parts of the Canadian Shield, so far as known, are devoid of Paleozoic sediments and made up entirely of pre-Cambrian rocks. They lie on either side of Hudson Bay; the western one is south of Coronation and Queen Maude gulfs and the eastern south of Hudson Strait (fig. 1). Hudson Bay is probably underlain by nearly flat-lying early Paleozoic strata. A vast region to the north in the Arctic Archipelago is made up of both elements, namely, pre-Cambrian rock and flat-lying early Paleozoic beds. If the new Geologic Map of the Dominion of Canada or the Geologic Map of North America is examined, the areas of pre-Cambrian and Paleozoic rocks are seen to be large but irregularly distributed. They include Victoria, Prince of Wales, Somerset, Baffin and Devon islands, and Boothia and Melville peninsulas. In some places the contact of the Paleozoic beds with the pre-Cambrian is exposed, in others the base of the Paleozoic strata is below sea level, and cliffs eroded in the horizontal strata rise precipitously from the shoreline to a height of 1,000 feet. On Baffin Island just north of Cumberland Sound the pre-Cambrian rocks are reported to rise 10,000 feet above sea level. In places Cambrian beds, elsewhere Ordovician and Silurian strata, rest directly on the pre-Cambrian.

It has been pointed out (Armstrong, 1947, p. 320) that, in general, successively younger systems of rocks make their appearance from south to north. On Ellesmere Island there may possibly be a continuous sequence from Cambrian to Triassic or even Jurassic. The Paleozoic rocks also generally increase in thickness from south to north. They range, presumably, from 100 or 200 to 1,000 feet in many coastal areas, but on

Borden Peninsula at the north end of Baffin Island there are about 2,000 feet of Ordovician and Silurian beds. Still farther north in the area of Carboniferous rocks on Ellesmere Island the total Paleozoic section is at least 10,000 feet thick. The Carboniferous rocks on the south side of Melville Island are reported to be 4,000 feet thick. These least two occurrences are mentioned in the next section.

The early Paleozoic strata are dominantly limestones and shales, with sandstone units generally in the Cambrian and Carboniferous systems. This is typically a central stable region or shield assemblage. The overlaps on the pre-Cambrian and the thicknesses also reflect shield geology.

It is tempting to try to define basins and domes or arches in the Paleozoic sediments of the Archipelago. An arch of pre-Cambrian rock may be seen stretching northward along Boothia Peninsula into Peel Sound, and West of the apparent arch is a Paleozoic basin that occupies the southern part of Victoria Island. Elsewhere, however, many scattered remnants of Paleozoic on the pre-Cambrian defy grouping into basins. In none of the areas do the reported thicknesses indicate local basins. If it be concluded that much of the pre-Cambrian was covered with early Paleozoic deposits and that, because of unequal upward movements since, they have been dissected and removed over large areas, it is a mistake to use the present base of the Paleozoic as a guide to Paleozoic basins. In the present state of information, however, it seems best to rest the case on the premise that only a broad division known as "pre-Cambrian and Paleozoic areas" can be recognized. It is surely an integral part of the Canadian Shield.

CARBONIFEROUS AND TRIASSIC AREAS

North and West of the area of pre-Cambrian and early Paleozoic rocks is a broad elongate area of Carboniferous strata. They may be followed from Robeson Channel between the northern ends of Greenland and Ellesmere Island southwestward to Dobbin Bay, then westward for about 100 miles, where a large Triassic area is found. To the south on Bjorne Peninsula more Carboniferous rock is reported; then southwestward to Grinnell Peninsula, which is also apparently made up entirely of rocks of the same age. West of this, Bathurst, Melville, Prince Patrick islands, and the northern part of Banks Island are probably underlain mostly, if not entirely, by strata of Carboniferous age.

A large area of Triassic lies west of the Carboniferous on the west coast of Ellesmere Island and the east coast of Axel Heiberg Island. It seems to have a linear nature and is parallel to the general Carboniferous boundary just described. This is especially conspicuous if the small areas of Triassic on Bathurst and Prince Patrick islands are joined with the main mass. The Triassic rocks are sandstones, with "subordinate schists and limestones" (Armstrong, 1947, p. 320). Some of the beds from Scoresby Bay to Cape Cresswell along the east side of the north end of Ellesmere Island may be Triassic, although no fossils have been found in them. They are "slates, quartzites, schists, grits, and limestones" and are known in northwest Greenland as the "Cape Rawson beds."

ELLESMERE-GREENLAND FOLDED BELT

Most of the rocks of Paleozoic and Triassic age in the Arctic Archipelago are flat-lying, but in a northeast-southwest belt through Ellesmere Island and the northwest coast of Greenland the beds

are both conspicuously folded and thickened. According to Armstrong (1947, p. 322):

> . . . In the vicinity of Vendome Fiord, on the west coast of Ellesmere Island, folded strata of Silurian and Ordovician age have been observed. A series of northeast-trending, sharp anticlines occurs in this region. Towards the south the flexures become less acute, although in many places, as along the north coast of Baumann Fiord, the strata have a very steep dip. Folding also becomes less severe towards the east and is barely recognizable 11 miles from the head of Makinson Inlet. Little is known of the extent of the folding in other directions. No folds can be seen on the south side of Baumann Fiord where the folded Silurian and Ordovician strata are either buried beneath flat-lying, younger Devonian and Carboniferous strata or swing to the west beneath the sea. If the former is the case the folding is pre-Devonian and probably of late Silurian (Caledonian) age.
>
> Another belt of folded rocks is exposed along the northeast coast of Ellesmere Island north from Scoresby Bay to Cape Cresswell. These folded rocks, called Cape Rawson beds, consist of unfossiliferous slates, quartzites, schists, grits, and limestones. The beds are commonly vertical and have a general southwest trend. Their age is not known. Bentham suggests they are possibly the northeast continuation of the folded Silurian strata on Vendome Fiord, and that the folding at both places is of Caledonian age. Koch, who has studied the Cape Rawson beds in northwest Greenland, also believes the folding is of Caledonian age, though his reasons are not conclusive. Schei suggests that the folding is of post-Triassic and pre-Miocene age, and that some of the Cape Rawson beds are Triassic. He bases this conclusion on the fact that he found lithologically similar folded Triassic rocks on the west side of Ellesmere Island.
>
> Triassic sandstones with subordinate schists and limestones underlie most of the east coast of Axel Heiberg Island and the opposite coast of Ellesmere Island. These rocks have provided diagnostic fossils. Schei states that the dip of the Triassic strata is in many places 50 to 60 degrees, and that folds occur north of Greely Fiord.

The Carboniferous rocks along the south coast of Melville Island near Bridgeport Inlet are known as the "sandstone series" and are at least 3,850 feet thick. They consist of 1,225 feet of shale, with a few sandstone beds, which are overlain by 2,625 feet of white sandstone. The beds have an average dip of 65° south (Armstrong, 1947, p. 319). The coincidence of strike of these beds with the general contact of the Carboniferous rocks and the older strata and the substantial thickness at the locality of high dips suggest that the Ellesmere Island folded belt curves westward through Melville Island. Since very little is known of the geology of Banks Island, the extent west of Melville Island of the folded belt is not known. It is possible that it lies beneath the Beaufort Sea and is expressed by northeast-trending Paleozoic beds in the Tanana-Yukon region of Alaska that Mertie (1935, p. 300) has described.

Because the Triassic beds are folded, it follows that the deformation is either post-Triassic (which would probably be Laramide, judging from Alaska) or in part post-Triassic and in part post-Carboniferous and pre-Triassic. If the latter view is correct, the Triassic beds should be in part coarse, owing to the earlier diastrophism. They are described only as "sandstones" and "schists." The latter word is evidently not used in the American sense, and its meaning is not clear. Possibly the Carboniferous beds overlie the Devonian and older beds unconformably, or possibly an angular unconformity separates the Devonian from the Silurian, and both the Acadian and Caledonian orogenies are represented in the compressional belt. These relations are only conjectural. The coincidence of the belts of Carboniferous rocks and diabase-intruded Triassic with the belt of folding suggests an Appalachian age, at least, for the first main phase of deformation, with

a strong post-Triassic phase definitely indicated.

TERTIARY DEPOSITS

Outcrops of Tertiary sedimentary rocks with coal beds occur at numerous localities throughout the Archipelago. At the north end of Baffin Island on the south side of Eclipse Sound, on both sides of Navy Board Inlet, and at the head of Isabella Bay interbedded sandstone, shale, and lignite, presumably of Tertiary age, occur. Deposits containing fossil wood crop out in the southwestern part of Prince Patrick Island. Tertiary deposits occur at several places along the northeast and west coasts of Ellesmere Island. They consist of sandstone, shale, and lignite and contain fossil plants. One seam of coal on Stenkuls Fiord is 5 feet thick. Tertiary deposits containing carbonized fossil tree trunks have been found in deposits on the northwest coast of Banks Island.

A Miocene age of the Tertiary deposits is mentioned by Armstrong (1947, pp. 316–324), from whose writing the above résumé is taken. Apparently, Chaney (1940, p. 480) charts them all as Eocene. The Tertiary deposits of the Arctic Archipelago are correlated with the Tertiary of the north coast of Alaska and especially with the "coal formation" of the interior of Alaska, generally thought to be Eocene, although it may be younger. Two coal-bearing series may be present in Alaska, and the same may be true in the Archipelago, but the general Tertiary age of the lignite-bearing strata of the Arctic Archipelago can hardly be doubted. The deposits are widespread and undoubtedly indicate a temperate, moist climate with extensive fresh-water swamps at elevations near sea level. If as young as Miocene, the great climatic change has come to the Arctic in late Cenozoic time. This is discussed under a later heading.

PLEISTOCENE EPEIROGENY

Washburn (1947) describes well-preserved, raised strand lines and marine fossils, which show that Victoria Island has emerged at least 500 feet. In addition, he believes that the whole of the Arctic Archipelago has undergone comparable uplift. This movement is interpreted as an isostatic adjustment to the unloading of Pleistocene ice. Elevated beaches up to 1,000 feet above sea level occur along the east coast of Hudson Bay (G. M. Stanley, personal communication).

ARCTIC SEA BASIN

SHELF AREAS

Soundings in the Arctic Sea areas shown on the Hydrographic Office Chart No. 2560 of the U.S. Navy Department do not include those of the U.S.S.R. North Pole Expedition under the command of Ivan Papanin in 1937 and 1938, and the drift of the "S.S. Sedov" through the polar sea in 1938 and 1939. These Soviet soundings plotted on the U.S. Navy chart, together with others from the North Atlantic and North Pacific, made possible the bathymetric contours of figure 1. A broad shelf area north of Siberia, less than 200 meters deep, is particularly prominent. Much of the sea from Nova Semlya (Novaya Zemlya) and northern Norway to Spitzbergen is less than 200 meters deep and is nowhere as much as 1,000 meters. No deep water has been reported in Hudson Bay or in the Arctic Archipelago except between Baffin Land and Greenland.

DEEP BASIN

With so much of the area of the Arctic Sea a shallow shelf, the truly deep basin

is not so large as commonly supposed. It is about 1,100 miles wide and 2,200 miles long. The Gulf of Mexico where deeper than 2,000 meters is 300 miles wide and 900 miles long, and the Caribbean Sea is about 400 miles wide and 1,600 miles long. The combined area of the deep basins of the Gulf of Mexico and the Caribbean is somewhat less than that of the deep basin of the Arctic Sea but, as far as size is concerned, the Arctic Sea is much more a mediterranean than an ocean. The Gulf of Mexico and the Mediterranean Sea are considered to be within the framework of the continents and not true oceans, but the Caribbean is thought by some to be bottomed by Pacific Ocean crust (Schuchert, 1935, p. 35).

The few deep soundings that have been made in the Arctic Sea reveal depths of from 3,000 to 6,000(?) meters, with much of the bottom about 4,000 meters deep. Detailed topography is suggested, but the few soundings could be contoured in a number of ways, depending on the interpretation desired.

PLATFORMS, RIDGES, AND BASINS

The great shelf north of Siberia extends to Alaska and southward beyond the Bering Straits into the Bering Sea (fig. 1). A region connecting Siberia with North America as broad as Alaska is thus under less than 200 meters of water.

The Atlantic Platform, or Mid-Atlantic Rise, extends northward to Iceland, where a transverse, narrow platform stretches from Great Britain to Iceland and to Greenland. Another rather narrow ridge or platform, not anywhere over 1,000 meters deep, connects northern Greenland with Spitzbergen and Spitzbergen with Norway.

There is a great basin on both sides of Greenland. One between Greenland, Spitzbergen, Norway, the Faeroes, and Iceland is irregular but deep, in one place over 4,000 meters. It is about 500 miles wide and 1,300 long. The other basin, that of Baffin Bay on the west side of Greenland, is slightly over 2,000 meters deep over much of its bottom, about 350 miles wide, and about 650 miles long. It is separated by a saddle, a small basin, and another saddle from the main Atlantic, toward which the waters progressively, but irregularly, deepen. The broad saddle area is about 1,000 meters deep.

GEOLOGY OF LANDS SURROUNDING ARCTIC SEA

SHIELDS FACING BASINS

Canadian shield.—The major tectonic elements around the Arctic Sea are the shields. Aside from the Alaskan orogenic belts and their continuation into the Anadir Peninsula of Siberia, the Arctic Basin is surrounded by the great stable elements of the earth's crust. These shields are separated from each other, or perhaps broken, by intra-shield orogenic belts which project into the Arctic Sea and are covered by it.

The Canadian shield, as explained above, extends to the orogenic belt of Ellesmere Island in the Arctic Archipelago. Beyond the orogenic belt is the undisturbed Carboniferous and Triassic Basin, the beds of which dip gently under the Arctic Sea.

Greenland shield.—Around the coast of Greenland are extensive areas of pre-Cambrian rock, and, except for the belt of Caledonian folding along the east side and the late Paleozoic folding along the north, the great island seems to be very similar to the Canadian shield, if not a part of it (Koch, 1929; and Hobbs, 1932, p. 373). Baffin Bay is a fairly deep depression between the two (fig. 1) and, be-

cause of its closed nature, must be of structural origin. This aspect of the geology will be dealt with later.

Russian-Baltic shield.—As depicted by Umbgrove (1947, pl. 5), the Russian-Baltic shield extends from the Ural Mountains of the U.S.S.R. westward to the Caledonian orogenic belt of the British Isles, Norway, and Spitzbergen. It, like the Canadian shield, consists of pre-Cambrian rock veneered in part by Paleozoic strata, little deformed except for basins of subsidence. North of the shield is a "massiv" or "nucleus," according to Umbgrove's plates 1 and 2, that has the tectonic character of a small shield. It is called the "Nucleus of Barents Sea and Putkow Kamen."

Angara shield.—The great Angara shield of Siberia extends from the Urals to the orogenic belt just east of the Lena River. Probably it was connected with the Russian-Baltic shield until Carboniferous time, when the Ural Mountains were formed in an intra-shield orogenic belt. The eastern edge of the shield is flanked by both Carboniferous and late Mesozoic orogenic belts.

An arcuate Caledonian and Variscan orogenic belt through the Taimir Peninsula and North Land cuts off a piece of the shield and, together with the Variscan belt through Nova Semlya, effectively surrounds the fragment. Umbgrove has called it the "Nucleus of Kara Sea."

Nucleus of Tschuktschen.—East of the Lena River is a broad belt of poorly known northward- and northwestward-trending structures which Umbgrove shows to be composed of Variscan and Mesozoic elements. They appear to continue northward through the New Siberian Islands into the Arctic Sea. The eastern part of the New Siberian Islands is pre-Cambrian rock, and so also is Wrangel Island. These exposures, together with the broad shelf, led Umbgrove to depict another nucleus, namely, the "Tschuktschen" flanking a large part of one side of the deep basin of the Arctic Sea.

PALEOZOIC OROGENIC BELTS

Most of the Paleozoic orogenic belts that occur in the Arctic region have already been mentioned. All are shown on the map of figure 1. The Caledonian belt through Great Britain, Norway, and Spitzbergen is well known. Folds and thrusts along the north and east sides of Greenland are thought to have Caledonian characteristics similar to those of Norway and Great Britain (Koch, 1929, 1935, 1939). The beds are overthrust westward along the east side of Greenland.

The age of the folds of the North Greenland–Ellesmere Island belt is uncertain: they might be Caledonian, Variscan, or even Mesozoic (p. 423).

The Variscan orogenic belt of the Ural Mountains splits before reaching the coast, the right arm extending through Nova Semlya into the Arctic Sea and the left arm northwestward to the coast and under the Barents Sea.

The orogenic belt along the east side of the Angara shield also extends to the coast and runs under the Arctic Sea. Variscan elements are believed to extend along the north shore of the Anadir Peninsula and to curve northward through the New Siberian Islands, where Caledonian folds also have been noted (Umbgrove, 1947, pl. 2). The two extend under the sea. It is possible that the Variscan folds find a continuation in Alaska, but, if so, the evidence is mostly covered and very incomplete.

The possibility of extending the Paleozoic orogenic belts under the Arctic Sea

FIG. 7.—Major lands and areas of deposition during the Carboniferous around the Arctic Sea. Data Koch (1939), Gignoux (1943), Stamp (1947), and Umbgrove (1947).

Sea. Data outside North America compiled from

and connecting them so as to form some kind of logical tectonic pattern is intriguing, but probably several patterns could be devised, and all must be considered a matter of guess. There is no doubt, however, that they project into the Arctic Basin and under the sea. They represent belts of weakness or mobility generally within or between shields and, together with the shields, form a single great tectonic province. It is difficult to escape the conclusion that the shelves, platforms, ridges, basins, and even the deep basin of the Arctic Sea are all part of the province and, more fundamentally, are continental—not oceanic—crustal material.

The great Paleozoic geosyncline and volcanic orogenic belt of western North America and Alaska is of the Pacific border province and distinct from the intra-shield orogenic belts. Alaska is the only approach of the great Pacific belt to the Arctic Sea.

CARBONIFEROUS BASINS

An attempt is made in figure 7 to show the great lands and the areas of deposition around the Arctic Sea in Carboniferous time. For the American Arctic the new Geologic Map of North America and the new Geologic Map of Canada and the accompanying publication, Economic Geology Series No. 1, were consulted; for Greenland, Koch's (1929, 1935, 1939) various publications; for Spitzbergen, the recent summary article by Allan (1942); for Great Britain, the recent book by Stamp (1947); and for Eurasia, the treatises of Gignoux (1943) and Umbgrove (1947).

A somewhat different map appears when the Carboniferous areas of deposition and the lands from which the sediments came are plotted (fig. 7) from the map showing orogenic belts and shields (fig. 1). In places the earlier Paleozoic orogenic belts had become highlands adjacent to the areas of deposition in Carboniferous time. The Scandinavian-Scottish "continent" seems established (Stamp, 1947, pp. 114–118), but its western extent and relation to the Appalachian geosyncline is unknown. The southern extent of the Carboniferous basin on the east coast of Greenland is a matter of speculation. The southward limit of the Carboniferous basin in the Arctic Archipelago of North America is clear cut, and it need not be projected far eastward to meet the Spitzbergen and eastern Greenland basins or far westward to meet the Cordilleran geosyncline.

Northeastern Siberia is so poorly known that several interpretations of its major tectonic features have been made. The commonest one is that it was all land in Carboniferous time; but this is incongruous with the geology of Alaska, where the great Cordilleran geosyncline existed. From the 1937 geologic map of the U.S.S.R., compiled and printed by the Soviets, it is surmised that the geology is somewhat similar to that of Alaska. This, together with Umbgrove's conception of the "Nucleus of Tschuktschen," has led me to show Carboniferous deposition along northeastern Siberia as well as land to the north. The land seems necessary to supply the mainland assemblage of sediments to the Cordilleran geosyncline in Alaska. This postulated land could have connected with the Canadian shield and not have been cut from it by the Archipelago basin as shown on the map.

The Russian Sea may not have branched into three lanes as shown but may have enveloped the whole broad terrane. At any rate, it appears that the lands and the areas of sedimentation of the northern hemisphere in Carboniferous times were a closely knit province

about the Arctic region and that they stood in contrast to the great Pacific Ocean and its marginal geosyncline and orogenic belts. The Appalachian geosyncline, also a marginal type according to Kay (1947, p. 1290), seems anomalous in the setting if it is projected northerly into the shield province. Its Carboniferous elements have generally been connected with the Armorican belt of orogeny in southern England and Brittany of France, but with great uncertainty.

MESOZOIC BASINS

Triassic sedimentary rocks north of the folded belt in the Arctic Archipelago have already been mentioned. They dip gently under the waters of the Arctic Sea; their thickness is unknown. When charted on a map of the Arctic region (upper map of fig. 8) they appear to be part of the same basin as the Triassic and Jurassic rocks of the Alaskan Arctic coastal plain north of the Brooks Range. Numerous areas of Triassic and Jurassic rock are shown on the U.S.S.R. Geologic Map of 1937 in the Anadir Peninsula of Siberia across the Bering Straits from Alaska, and presumably they represent the continuation of the same great basin. The Mesozoic rocks continue westward in Siberia to the Lena River and Angara shield and northward through the New Siberian Islands. The distribution shown on the map of figure 8 is based on the work of Arkhanguelsky (1937, p. 299).

Additional Mesozoic sediments occur extensively on the east coast of Greenland (Koch, 1929) and on Spitzbergen (Allan, 1942) (fig. 8). The Jurassic sections in these areas have representatives of most of the epochs. Northern Norway also has Oxfordian and Portlandian strata of the Upper Jurassic. The seas of Callovian time set their marks on Franz Joseph Land and Nova Semlya (Koch, 1929).

Paleogeographic maps of the Arctic can be regarded only as hypothetical. The interpretations of the Siberian Arctic are especially weak. Although our knowledge of Alaska is incomplete, it is sufficient to bring out the incongruous and probably erroneous nature of certain maps that have been published on Siberia across the straits. The only value of an attempt to make a paleogeographic map of the Arctic is to express the belief that the region was no different from other parts of the continents of the Northern Hemisphere. As for the Triassic and Jurassic seas, they transgressed the great shields extensively, and certain lands existed where water now lies. The physical evidence is not enough to postulate land connections between the Eurasian and North American continents in the Greenland-Norway region, but they could have existed at times. The Alaskan connection appears to have been almost continuous because of the geosynclinal and orogenic belt there.

The seas of Cretaceous time were more complex than those of the Jurassic and Triassic, and an attempt to assemble reliable data and to make a map of the seaways led only to the conclusion that it was more a guess than an interpretation, and therefore not worth while. It need only be said that marine Cretaceous strata occur widely and in unexpected places.

TERTIARY DEPOSITS

Numerous Tertiary deposits have been found in the Arctic region, and fortunately most of them carry coal beds and plant fossils. The lower map of figure 8 is taken partly from a study of the plants by Chaney (1940, pp. 469–488). He attempted to reconstruct the character and

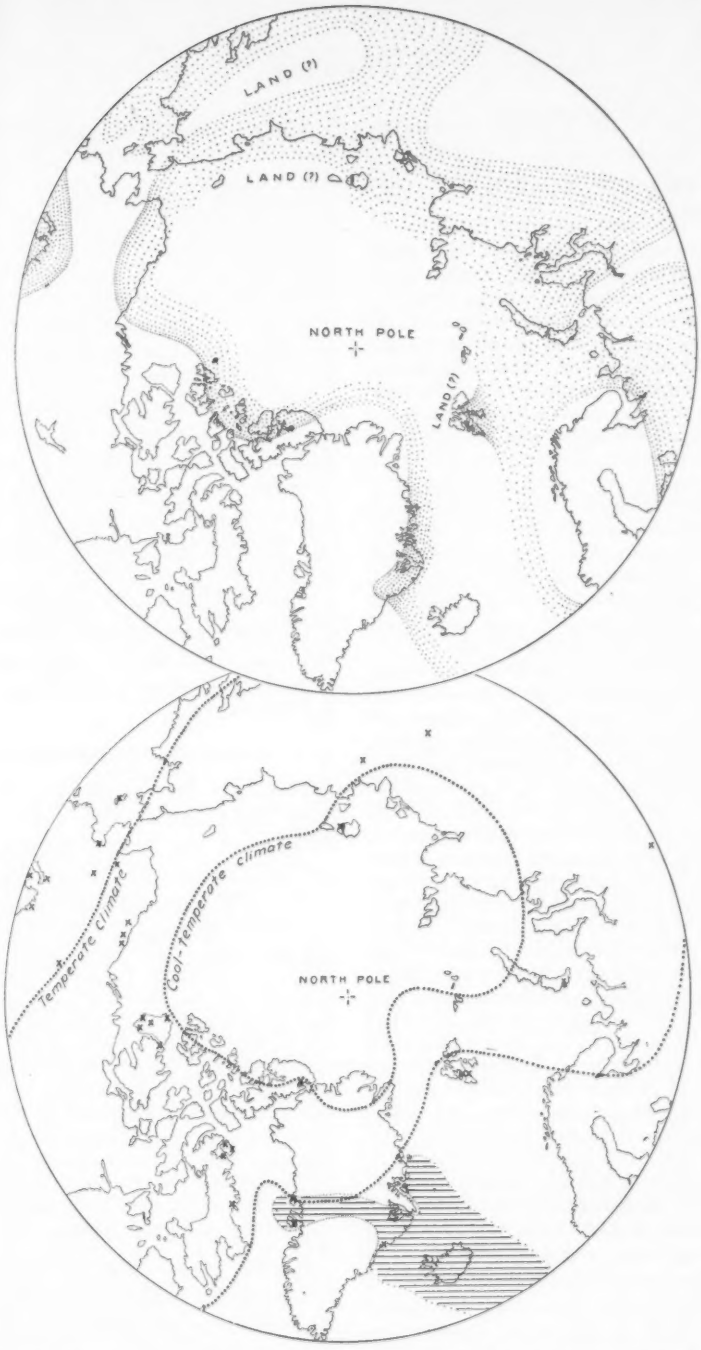


FIG. 8.—*Upper map:* very generalized distribution of seas and lands of the Arctic during Triassic and Jurassic times. The seas at any one time were not so extensive as the total distribution shown.

Lower map: early Tertiary deposits of the Arctic. The dotted lines are isoflors after Chaney (1940), and the crosses denote Chaney's Eocene and Oligocene localities, plus a few other localities where "Arctic Miocene" coal beds are known. The ruled area denotes the Greenland-Iceland-Scotland basalt field of early Tertiary time.

distribution of the Tertiary vegetation, in order to learn of the distribution of climate and migration routes in the past. He writes:

> The Tertiary flora which best illustrates a migration response to climatic change has as its dominant element the redwood (*Sequoia*), together with several of its modern associates, alder (*Alnus*), pepperwood (*Umbellularia*), and tan-oak (*Lithocarpus*). An associated element includes members of the modern deciduous hardwood forest of the eastern United States and eastern Asia, basswood (*Tilia*), beech (*Fagus*), chestnut (*Castanea*), elm (*Ulmus*), and hornbeam (*Carpinus*). A third element is made up of general now restricted to Asia, such as katsura (*Cercidiphyllum*), maidenhair tree (*Gingko*), tree of heaven (*Ailanthus*), water chestnut (*Trapa*), and zelkoua (*Zelkova*). Such plants are typically temperate in their modern distribution. Their moisture requirements are high (40 to 60 inches annually), and the predominance of deciduous genera suggests summer as well as winter rainfall. This flora first appeared in the Eocene of high northern latitudes, including Alaska (Kenai flora), (Hollick, 1936) northern Siberia, Spitzbergen, and Greenland. Evidence that some of its genera ranged southward along mountain ranges during this epoch is to be seen in its occasional presence as a minor element in the subtropical floras which characterized middle latitudes in western America and western Europe during the Eocene. The existing forests of Alaska include a number of trees, such as the mountain hemlock (*Tsuga mertensiana*) and aspen (*Populus termuloides*), whose modern distribution southward at successively higher altitudes in the Cordillera corresponds closely to that here suggested for the Eocene redwood forest. Except in the northern portions of their range, where they live at sea level, both of these trees occupy habitats so remote from sites of deposition that their entrance into the contemporary record is rather unlikely. If a future change in climate should lower the temperature and precipitation in the western United States, the forests now living in our lowlands would give way to vegetation from the north and from the Cordillera. This northern and alpine vegetation would leave a record as soon as its altered range brought it down into the valleys, in close proximity to sites of deposition.

Chaney's conclusions are that the climatic zones were distributed around the North Pole in early Tertiary time with modifications due to land and warm ocean currents, as today, except that the rain-forest climate (temperate) of the California Coast Ranges was near the pole, and subtropical conditions existed in California, or as far north as latitude 40; that the factors controlling air and water circulation were essentially the same then as now. He believes that the gradual cooling of the Arctic region is coincident with a gradual uplift of the continent and that there was a land connection between Siberia and Alaska almost continuously during the Tertiary.

Figure 8 also shows the extent of the great basalt field between Greenland and Scotland. It is possible that much land existed in this area during the eruptions and afterward.

TERTIARY CRUSTAL MOVEMENTS

Following the deposition of the Tertiary coal-bearing beds, the region of Spitzbergen was gently uplifted in the eastern and central districts, but steep folding and overthrusting occurred along the west. The large island was taking its modern shape, and, according to Allan (1942, p. 46),

> Stor Fjord was a major depressed fjord block and Ice Fjord with its branches was probably due to a combination of east-west tear-faulting and associated block faulting. Today the etched and buttressed slopes of the horizontally bedded rock masses flanking Ice Fjord display numerous examples of step faulting toward the shore lines. The Tertiaries on either side of the Foreland Sound dip toward it—yet another trough faulted feature.

Rather gentle folding occurred in Britain in the Miocene and reflected the superior mountain building in the Alpine chains of the continent, and by the close of the Miocene the structure was vir-

tually complete. At the beginning of Pliocene time southeastern England was 650 and 700 feet lower than today (Stamp, 1947), and, while the land rose, a succession of platforms was carved by the waves. One of the best is the "400 foot platform"—400 feet above sea level.

Faults have been recognized on both the west and the east coasts of Greenland. The Cape York district of northwest Greenland is especially broken by high-angle faults (Koch, 1929), and the fiords of the west coast about Disko and Umnak bays generally have their courses parallel to faults (Hobbs, 1932, p. 380). It is not clear, however, that these faults were associated with Tertiary land movements. Koch (1935) believes that strong Tertiary faulting may be recognized in many places along the eastern coast and that it was associated with the great volcanic activity. The faults have tilted a plane to the west on Milne Land and may be seen cutting the sediments there. Along the east side of Hurry Inlet are Tertiary faults, and Liverpool Land was doubtless strongly raised in Tertiary time. On Wegner peninsula and in nearby fiords, on the east side of Traill Island, on Geographical Society Island, on Ymer Island, and several others north to the Wollaston Foreland, many faults have been recognized and are believed to be of Tertiary age. Part of the movements predate the great lava flows, and part postdate them. The volcanics of eastern Greenland are several thousand meters thick in places and, as a number of writers have proposed, must be continuous with the basalt fields of Iceland, the Faeroes, and Scotland.

The geology of northern Siberia is still very obscure, but a large marine transgression from the north occurred in Quaternary time, and much of the geology of this vast region is hidden under the blanket of the sediments of this sea and also certain glacial deposits. The Khatanga depression along the Khatanga River, which cuts the northern part of the Angara shield, has been considered to be bounded by faults and to be of Tertiary age. According to Shanazarov (1948, p. 172):

> . . . Epeirogenic phenomena took place during the entire Tertiary and Quaternary, bringing about the sinking of some blocks of Siberian territory and the raising of others. According to recent observations these epeirogenic movements are continuing even to the present.

Although the Arctic localities where Tertiary faulting and land movements have been observed are relatively few and the details and ages are not precisely known, they do suggest crustal movements that could have affected profoundly the distribution of land and sea. It seems entirely possible that faulting and epeirogenic movements somewhat comparable to those of the Great Basin in western North America in mid- and late Tertiary time could have occurred and that the basins of Baffin Bay, the Greenland Sea, and even in part the deep Arctic Sea basin could have been formed in somewhat their present shape during the Tertiary. The great thickness of basic eruptives in the Scotland-Greenland region must have altered the land areas considerably there also.

PLEISTOCENE EPEIROGENY

Isostatic uplift due to the melting of the glaciers, of 600 feet in Victoria Island and of 1,000 feet in Hudson Bay, has already been mentioned. The Scandinavian Peninsula was depressed by the last ice and has risen 600 feet since its dissipation. In all places the rise is believed to be continuing.

If the ice now on Greenland were to melt, the land there would rise about 600

feet if the Scandinavian region is taken as a precedent. According to James W. Wilson's isostatic computations for the writer, the center of Greenland would rise 600–1,000 feet and the uplift would extend beyond the present shores but not enough to connect Greenland with Spitzbergen or Iceland. It probably would merge Ellesmere Island with Greenland (fig. 1).

If the fossil record should demand major land connections between Greenland and Europe during the Pleistocene, it is evident that movements other than those caused by loading and unloading by the ice must be sought. These seem to be a continuation of the late (?) Tertiary movements just described that undoubtedly are of deep-seated origin.

SUMMARY OF EVIDENCE OF ANCIENT ARCTICA

A review of Alaskan sediments has led to the belief that the Paleozoic Cordilleran geosyncline of western North America extended through Alaska and embraced the whole of the territory. The Pacific side of the geosyncline is marked by the volcanic-bearing assemblage. Such an arrangement requires an adjacent land on the north, where sea now exists. This observation was the first that directed attention to the theory that the Arctic Sea was not an ocean with a permanent basin but that the region had been a land area at times in the past and that Eurasia and North America were broadly connected.

The next observation that supports this concept of Ancient Arctica is the study of the topography of the surface below water in the Arctic region. Much of the sea floor is very shallow and the truly deep basin is only slightly larger than the combined deep-water basins of the Gulf of Mexico and the Caribbean. The parts that range in depth from the shallow shelves to the deep Arctic Sea floor are marked by basins, escarpments, and ridges; and it seems probable that, when bottom profiles are made with the sonic depth finder, that much more detail will come to light. The soundings made of the deep Arctic Basin are pitifully few, and it does not seem logical to conclude that the floor is as smooth as the contours of figure 1 indicate. The depths in excess of 12,000 feet might be due to faulting; and the depression, when better outlined, might have graben characteristics. The presence of considerable topographic detail in much of the surface now covered with water in the Arctic region suggests that the crust there is composed of continental material, and, therefore, it is believed that the North Polar waters should be classed as a mediterranean sea—not an ocean.

The third supporting observation is the fact that the shields of the Northern Hemisphere all face the Arctic Sea and, except for Alaska, surround it with a common geology. It looks as if the deep basin were only a sunken part of a single, great continent.

The shields are separated by several Paleozoic orogenic belts that project to the Arctic Sea and, although in full development at its margin, are lost under it. They surely project long distances into the region now covered by water.

A study of the areas of sedimentation in Paleozoic, Mesozoic, and Cenozoic time on the lands around the Arctic Sea shows clearly that the epeiric sea conditions, which prevailed there at a number of times, were very similar to those of the central stable region and the Canadian shield of North America. Certain of the great seas that invaded North America started in the Arctic and progressively spread southward. Numerous Tertiary

deposits containing coal are known in the Arctic lands. They attest temperate, moist conditions in the Arctic, very unlike those of today, even as late as the Oligocene. Land was evidently more extensive at the time that the coal swamps existed than now, because the beds are now mostly submerged.

A sixth important observation that bears on the continental makeup of the Arctic is the evidence of faulting and major epeirogenic movements in Tertiary time. The late (?) Tertiary thrust faulting on Spitzbergen, the numerous high-angle faults in the marginal areas of Greenland, Pliocene uplift in Britain, and the faulting in the Khatanga depression and elsewhere in northern Siberia all betray widespread crustal unrest in Cenozoic time throughout the Arctic. It is not difficult to visualize these movements as having been of the magnitude of the late and mid-Tertiary faulting in the Basin and Range Province of western North America. As Britain had acquired its present structure and form by the close of Miocene time and as Spitzbergen took its present outline during the post-Miocene (?) faulting, so also it appears that the topography of the Arctic Sea floor, together with that of Baffin Bay, the Greenland Sea, and the North Atlantic, acquired its modern form at about the same time.

A major lava field or volcanic province stretches from Scotland to Greenland but is now covered largely by water. The volcanic rocks, early Tertiary in age, are very thick in places and perhaps at times effectively bridged Europe with America.

As an eighth observation, the isostatic uplift in response to the melting of the continental glaciers has not yet run its course, and a little more land will yet appear because of the continued rise. If the ice of Greenland were to melt, a rise of about 600–1,000 feet would occur, which would drain the water between Greenland and Ellesmere Island but not connect Greenland with Spitzbergen or Iceland.

As a ninth and last observation, the frequent, widespread distribution of animals and plants common to both Eurasia and North America requires migration routes that were probably continuous across the Pacific and common across the Atlantic. It appears to the writer that the geology of the Arctic permits such migration and that the implied connections are much more natural and plausible than a concept of long, narrow land bridges across the Atlantic or the drift of the continents.

PALEONTOLOGIC SIGNIFICANCE OF ANCIENT ARCTICA

Much has been written about the similarities of Appalachian trilobites with those of Britain and the common varieties of Spitzbergen and North America brachiopods. The realization was startling long ago that Asiatic cephalopods had counterparts in western North America. These seem simply explained by the migration routes that the shifting epeiric seas and orogenic belts of Paleozoic and Mesozoic Arctica afforded. Because of the warm water that the fossils of the far north indicate, climatic barriers to migration could not have prevailed.

Simpson (1947, pp. 613–688) has recently reviewed the genera and families of mammals common to Eurasia and North America and has ably examined the significance of faunal resemblance. He concludes:

> Major faunal interchanges occurred in early Eocene, late Eocene, early Oligocene, late Miocene, middle and late Pliocene, and Pleistocene. There was little or no interchange in the

middle Eocene and middle and late Oligocene. Each interchange involved migration in both directions, but there was probably more movement from Eurasia to North America than in the opposite direction. All interchanges were selective, and they became increasingly limited from Eocene to Pleistocene. The most important selective influence was probably the relatively cold climate of the land connection. This connection was probably from Siberia to Alaska throughout the Cenezoic and was in almost continuous existence with important interruptions in parts of the Eocene and Oligocene and perhaps shorter interruptions at later times.

Simpson also points out that the Bering connection is adequate to explain all known distributions and that an Atlantic connection is not necessary but still not ruled out. He also cites an oral communication from Colbert to the effect that the migration route of Cretaceous dinosaurs from central Asia to North America was Pacific, not Atlantic.

The bearing of the fossil plants of the Tertiary on climate and land distribution has already been discussed.

It is evident that the mammalian fossil evidence points out more times of land connections than the physical geology yet reveals. The physical geologist has only the Tertiary coal-bearing beds of the northern regions and their deformation as time markers to guide him. The Tertiary beds have generally been called "Arctic" Miocene, but most of them are probably late Eocene, and the subsequent orogeny, such as on Spitzbergen, may be Alpine. When more complete study has been made of the Tertiary beds cropping out in the Arctic, they may prove to be of several ages, and then more can be said about the physical history. Conclusions that may be drawn in consideration of both the fossil and physical evidence are as follows:

1. The Alaskan-Siberian region is one of extensive, very shallow water today and has been the site of geosynclinal and orogenic activity of the Pacific marginal type from early Paleozoic times to the present and has been, therefore, a region of frequent island arcs, mountain chains, and broad arches. As such it would have served effectively and persistently as a migration route of both land and shallow-water faunas. Here the organic record corresponds well with the physical.

2. Migration routes in pre-Carboniferous time, especially, need not have been limited to the geosynclinal belt but could have followed the lands and the stable region of what is now the Arctic Sea and the shores of the epeiric seas that invaded it from time to time. The scant data now available suggest that the deep basin of the Arctic Sea began to form in Carboniferous time and that the principal land migration routes thereafter circumvented it.

3. Although, according to Simpson, an Atlantic "bridge" is not necessary, the physical evidence indicates that connections could have existed. Judging from the topography of the sea floor, two restricted routes were the most recent, one from Asia and Scandinavia via Spitzbergen and northern Greenland and one from Great Britain via Iceland and central Greenland. It is also not inconceivable that extensive land areas connecting Greenland with Eurasia existed.

PETROLEUM POSSIBILITIES

PREVIOUS RECOGNITION

The Arctic region has not been overlooked in the search for petroleum. In a recent article Pratt (1947, pp. 658–659) accords it a promising future on three counts: first, it is an area of widespread continental shelves; second, the Arctic Sea is the world's "fourth mediterranean" and is still unexplored, whereas some

of the world's richest reserves are clustered about the other three; and, third, surface evidences of petroleum are known there.

The extent of shallow sea in the Arctic can be determined approximately by inspection of the map of figure 1, and it is true that the amount is great. The concept that the Arctic Sea is a mediterranean has already been elaborated, and it is pleasing to note that Pratt arrives at the same conclusion. It is largely an intra-shield mediterranean, however, and not everywhere associated with Cenozoic orogenic belts, like the others.

The Arctic coastal plain and foothills to the Brooks Range are already being explored, and successful results are expected (Reed, 1946, pp. 1441–1443).

Greenland presents only remote oil possibilities because of its pre-Cambrian terrane, its Caledonian orogenic belts, its Tertiary lavas, and, finally, its great ice-cap.

Spitzbergen, Nova Semlya, North Land, and the New Siberian Islands are in large part Paleozoic orogenic belts and would normally not interest the petroleum geologist, although bituminous limestones and shales are reported on the New Siberian Islands (Shanazarov, 1948, p. 183). Fohs (1948, pp. 317–330) and Shanazarov (1948, pp. 153–197) have examined literature on northern Siberia for petroleum possibilities. Of the Arctic areas in Siberia the Khatanga depression seems the most promising, although largely unknown and difficult to explore. The mountains of the vast area of northeastern Siberia and the Anadir Peninsula are generally unknown geologically, but vast areas are underlain by marine Mesozoic deposits which at places seem to offer favorable structural conditions for the accumulation of petroleum (Shanazarov, 1948, p. 183).

ARCTIC ARCHIPELAGO

Very little attention has been accorded the Arctic Archipelago for its oil possibilities, although it seems to the writer that it is one of the most promising regions of the Arctic. North of the Greenland-Ellesmere folded belt there appears to be a foothill and coastal plain province of Carboniferous, Triassic, and, perhaps, Jurassic rocks, which extends through the northern part of Ellesmere Island, the Sverdrup Islands, Cape Grinnell Peninsula, Barthurst Island, Melville Island, Prince Patrick Island, Borden Island, and the northern part of Banks Island (examine maps, figs. 1 and 8, and the Geologic Map of North America, 1947). Although little explored, the strata are the usual epeiric sea sediments with some intercalated fresh-water coal beds. They are flat-lying or gently folded north of the folded belt. Actual oil seeps of petroleum or bitumen have been reported on northern Melville Island (Armstrong, 1947, p. 320). Such a region possesses the geologic requisites of an oil province. It is not known whether the seeps are from Triassic or Carboniferous strata, but, if from Triassic, the Carboniferous beds will be found at depth. Altogether, there are about 100,000 square miles of land in that part of the Archipelago underlain by Carboniferous strata, and they are worth careful investigation. The deterrent to detailed study is the Arctic climate; but, with airborne support, geologists could do a great deal of profitable field work in the summer months. In addition to hindering field work, the Arctic climate would make the delivery of oil by ship very difficult; but if the need were sufficiently great it seems possible that tankers could get in and out during August and September, at least, of each year. A study of the elaborate publication, "Ice Atlas of

the Northern Hemisphere" by the Hydrographic Office of the U.S. Navy Department, indicates that the unnavigable polar pack impinges on Prince Patrick Island, Borden Island, and the northern coast of Ellesmere Island the year round; but, in the months of August and September, Davis Strait, Baffin Bay, and Lancaster Sound are open, and a passage westward through Melville Sound to Beaufort Sea generally can be made by heavily built vessels without the aid of an icebreaker. It seems probable that contact with shore storage facilities could be made on any of the islands during at least two months of the summer, except along the shores facing the polar pack, if heavy icebreakers were used.

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SOME MELILITE SOLID SOLUTIONS[1]

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ABSTRACT

The compositional problems, especially the soda content, of the melilite group of minerals are reviewed. The system $Ca_2Al_2SiO_7$ (gehlenite)–$Na_2Si_2O_7$ was investigated; 15 per cent of $Na_2Si_2O_7$ was found to go into solid solution with gehlenite. Akermanite forms no solid solution with $Na_2Si_2O_7$. A portion of the system gehlenite–$CaO \cdot 2Al_2O_3$ was investigated to determine whether any solid solution takes place. Evidence, not decisive, suggests that gehlenite may take approximately 10 per cent of the calcium aluminate in solid solution.

INTRODUCTION

The melilite group of minerals presents compositional problems that several workers have attempted to resolve. The complete solid solution series between gehlenite ($Ca_2Al_2SiO_7$) and akermanite ($Ca_2MgSi_2O_7$) is now well known, and it is generally agreed that natural melilites are composed principally of mix-crystals of these two components. The iron analogue of akermanite, $Ca_2FeSi_2O_7$, has been synthesized (Bowen, Schairer, and Posnjak, 1933), and a zinc melilite, Hardystonite ($Ca_2ZnSi_2O_7$), is known. The iron and zinc found in melilites are thus easily explained. Many chemical analyses, however, show rather large amounts (up to a maximum of 6.11 per cent) of Na_2O and K_2O, and silica in excess of the values accounted for by mixtures of the akermanite-gehlenite "molecules."

A. N. Winchell (1924) took up the melilite problem from the viewpoint of volume isomorphism and largely discredited the sarcolite molecules which had been proposed earlier by W. T. Schaller (1916) and partially verified by A. F. Buddington (1922). Winchell pointed out the essential R_5O_7 character of the melilites and suggested $Na_2Si_3O_7$ and $Ca_3Si_2O_7$ as the most likely "molecules" entering the melilite structure. The soda and excess silica observed in some melilites would thus be accounted for, and excess lime (which Winchell thought existed in some melilites) would also be explained. In addition to the above hypothetical end-members, Winchell also suggested another means of explaining excess silica by supposing that SiO_2 is "interatomically" dispersed in the structure. This view was based on the concept, now known to be erroneous, that the oxygen atom was very small. Three more complex molecules were also considered as possible end-members; but there are structural objections to all of them, as in the case of interatomic SiO_2, and they will not be considered here.

Berman reviewed the problem in 1929 and modified Winchell's fundamental R_5O_7 to $X_2Y_3O_7$. He discarded all the end-members suggested by Winchell with the exception of $Na_2Si_3O_7$. Berman objected to the $Ca_3Si_2O_7$ "molecule" because of its 3:2:7 ratio (not found in natural melilites) and because, if this molecule is used as a basis for recalculation of melilites from the twenty-three chemical analyses he collected, there appears an apparent excess of as much as 9 per cent of silica. Recomputation of melilite compositions from the available analyses, using various end-members, led Berman to replace Winchell's calcium

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silicate with his "submelilite" molecule, $CaSi_3O_7$. The deficiency of $CaSi_3O_7$ in lime was used to account for the observation that many of the analyses do not show the theoretical Ca + Na total of 20, if oxygen is set at 70 (i.e., there is a deficiency in the X_2 value). Berman also interpreted the $CaSi_3O_7$ molecule as indicating that part of the melilite structure is open. The end-members postulated by Berman are:

Gehlenite —$Ca_2Al_2SiO_7$
Akermanite —$Ca_2MgSi_2O_7$
Soda melilite—$Na_2Si_3O_7$
Submelilite —$CaSi_3O_7$

Recalculation of the chemical data led Berman to state that in the natural occurrences the submelilite molecule does not exceed 10 per cent and the soda melilite does not exceed 25 per cent.

It is unlikely, however, that the submelilite "molecule" can go into solid solution in the melilites; and the same may be said for the $Ca_3Si_2O_7$ "molecule" of Winchell. Aside from difficulties inherent in the replacement of Ca for Al in gehlenite that would be required in the latter formulation, the work of Schairer and Osborn (1941) on the systems Gehlenite–$CaSiO_3$ and akermanite–$CaSiO_3$ disproves the existence of either of the two end-members in the melilites. Both systems are truly binary and show no solid solution. If *any* calcium silicate entered either akermanite or gehlenite, the above systems would show solid solution and would not be binary, as the Ca and Si would be removed from the melt in the proportions necessary to form mix-crystals, resulting in a nonbinary residuum. In addition to this rather conclusive evidence, it is doubtful that the melilite structure has the "holes" that would be required by Berman's formulation of $CaSi_3O_7$.

In 1930 Warren and Machatschki published papers on the melilites. Warren first pointed out, on the basis of X-ray evidence, that the general formula should be $(Ca, Na)_2(Mg, Al)_1(Si, Al)_2O_7$, as there is no evidence to indicate that Si and Mg can ever substitute for each other. Machatschki discussed the same point on the basis of the considerable size difference between Si and Mg and wrote the melilite formula as $X_2YZ_2(O, OH)_7$, where

$X = Ca, Na, (K)$,
$Y = Mg, Fe^{++}, Fe^{+++}, Al$,
$Z = Si, Al$.

This formulation, which doubtless is essentially correct, is the first in which the possibility of OH groups entering the structure is recognized. Machatschki also pointed out difficulties inherent in chemical analyses based on small samples and suggested that the small size and impurity of the samples might explain the low Ca + Na values often found. An alternative suggestion was that the larger cations (K, Na, Ca) might be leached out by secondary chemical action without greatly disturbing the structure; this, however, is unlikely in the case of the melilites, since removal of these ions would result in structural disruption.

THE SODA CONTENT OF MELILITES

THE SYSTEM $Ca_2Al_2SiO_7$ (GEHLENITE)–$Na_2Si_3O_7$

$Na_2Si_3O_7$, named "soda melilite" by Berman, does not exist as an independent compound. This fact, of course, does not preclude soda and silica from entering into solid solution as a component in the 1:3 ratio expressed by $Na_2Si_3O_7$. To determine the extent of soda substitution in gehlenite, the system $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$ was investigated.

The technique of investigation, devel-

oped at the Geophysical Laboratory, is now so well known that it will not be here described. Homogeneous glasses prepared from mixtures of pure SiO_2, Al_2O_3, Na_2CO_3, and $CaCO_3$ were subjected to thermal study by the quenching method. Figure 1 represents the results of this investigation. The thermal data from which the equilibrium diagram was constructed are listed in table 1.

$Na_2Si_3O_7$ goes into solid solution with gehlenite to the extent of 15 per cent by weight, equivalent to 3.85 per cent of Na_2O. The extent of the solid solution was determined by observation of the completely devitrified mixtures; nonhomogeneous crystallizations result when the amount of $Na_2Si_3O_7$ exceeds 15 per cent in the mix. The limit of solid solution can also be clearly determined with the aid of figure 2, which is a plot of index of refraction (Na vapor light) against percentage of $Na_2Si_3O_7$. Included on this plot are the indices of the glasses in the system. The indices of pure gehlenite ($\omega = 1.669$; $\epsilon = 1.658$) drop to values of $\omega = 1.644$; $\epsilon = 1.628$ at 15 per cent $Na_2Si_3O_7$, at which point the limiting index values expressed by a break in the curve indicate the limit of solid solution.

Inasmuch as the composition $Na_2Si_3O_7$ does not crystallize as an independent compound but forms SiO_2 and $Na_2Si_2O_5$ (Kracek, 1939), the system ceases to be binary at solidus temperatures beyond the solid-solution limit. The diagram was for this reason not extended beyond this point.

THE SODA-AKERMANITE AND GENERAL SODA-MELILITE RELATIONS

To obtain additional evidence on the presence of soda in the melilites, the relation between akermanite and $Na_2Si_3O_7$ was experimentally investigated. It was found that akermanite takes no $Na_2Si_3O_7$ in solid solution. This statement is based on the results obtained from one mix, of the composition 10 per cent $Na_2Si_3O_7$–90 per cent $Ca_2MgSi_2O_7$. This mix, if devitrified completely (i.e., below the solidus), consists of crystals of pure akermanite and a fine-grained residual crystalline aggregate, undoubtedly composed of SiO_2 and $Na_2Si_2O_5$, although no attempt

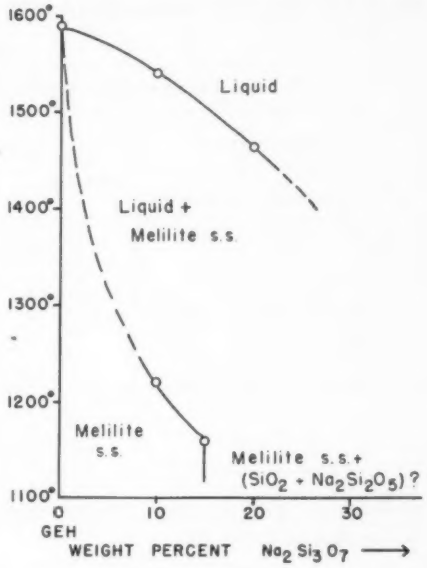

FIG. 1.—Equilibrium diagram for the system $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$.

was made positively to identify the constituents present. The inhomogeneous crystallization indicates that, if any solid solution exists, it is composed of considerably less than 10 per cent by weight of $Na_2Si_3O_7$; also, the refractive indices of the akermanite were the same as those of the pure synthetic crystals ($\omega = 1.633$; $\epsilon = 1.639$). The fact that the indices were not changed is very strong evidence that no solid solution was formed at all.

The conclusion that akermanite takes no $Na_2Si_3O_7$ in solid solution is not sur-

prising; in fact, it is to be expected. If the formulae be written

$Ca_2MgSi_2O_7$ (akermanite)
$Na_2Si\ Si_2O_7$,

it is seen that to form mix-crystals a substitution of Si for Mg would be required. The improbability of this substitution has already been mentioned; not only is the Mg ion considerably larger than the Si ion, but there is a valence difference of two charges. This situation does not exist in the case of gehlenite:

$Ca_2Al_2SiO_7$ (gehlenite)
$Na_2Si_2SiO_7$.

The Na-Ca and subsequent Si-Al substitution is well known in many minerals.

The solid solution relations are thus such that one end-member of a completely isomorphous series accepts a limited amount of a third ingredient, whereas the other end-member does not. Mixtures of the three components might then be expected to show a drop in the acceptance of the $Na_2Si_3O_7$ as the composition changes in the direction of enrichment in akermanite.

A mix composed of 45 per cent gehlenite, 45 per cent akermanite, and 10 per cent $Na_2Si_3O_7$ was prepared. The completely devitrified preparation consisted of melilite crystals plus a few per cent of inhomogeneous material, showing that less than 10 per cent $Na_2Si_3O_7$ is taken into solid solution at the mid-point of the gehlenite-akermanite series. The optical properties of the melilite, in the preparation described above, confirm this observation; the indices of crystals composed of equal amounts of gehlenite and akermanite are $\omega = 1.653$ and $\epsilon = 1.652$ (Ferguson, Buddington, 1920), the indices of the melilite formed from the 45–45–10 per cent mixture are $\omega = 1.640$, ϵ approx. $= 1.637$. Figure 2 shows that in gehlenite, 10 per cent $Na_2Si_3O_7$ causes a drop in mean index of approximately 0.020. The drop in the above solid solution is but 0.013—additional evidence that less than 10 per cent of the soda "molecule" enters this composition.

If the amount of $Na_2Si_3O_7$ varies in a linear fashion with the akermanite-gehlenite ratio, the mid-point in the isomorphous series should contain $7\frac{1}{2}$ per cent $Na_2Si_3O_7$. The amount of residual material in the one composition exam-

TABLE 1

THERMAL DATA FOR THE SYSTEM $Ca_2Al_2SiO_7$–$Na_2Si_3O_7$

Weight Per Cent		Time (Minutes)	Temperature (° C)	Final Condition
$Ca_2Al_2SiO_7$	$Na_2Si_3O_7$			
90	10	30	1,543	All glass
		30	1,539	Sparse melilite
		120	1,224	Melilite, traces of interstitial glass
		120	1,216	All melilite
85	15	150	1,165	Melilite, traces of glass
		120	1,155	All melilite
$82\frac{1}{2}$	$17\frac{1}{2}$	.....	.....	Nonhomogeneous crystallizations at subsolidus temperatures
80	20	.....	.....	Nonhomogeneous crystallizations at subsolidus temperatures
		30	1,464	Sparse melilite
		30	1,460	All glass

ined, as well as the refractive index of the melilite, agrees well with this value of 7½ per cent. Figure 3 is constructed from the above data and shows the limited extent to which soda can enter the structure of the synthetic melilites. Additional points, to fix with more certainty the position of the boundary delimiting the soda-bearing melilites, are not necessary. It is apparent that the soda solid solution is at a maximum in pure gehlenite and that the amount of soda in the synthetic melilites drops off in a linear fashion as the akermanite content of the mix-crystals increases, until reaching zero in pure akermanite.

Soda entering the melilite structure must do so at the expense of Ca, and, in so doing, either the Mg, the Al, or the Si must be replaced by an ion of higher charge. Thus, if a sodium ion replaces a calcium ion, one of the following associated anion replacements is possible; Si^{4+} for Al^{3+}, P^{5+} for Si^{4+}, or Al^{3+} (Ga^{3+}) for Mg^{2+}. In addition to the anion substitutions, cation substitutions of a different nature must be considered. If Na replaces Ca, electroneutrality can also be restored by the replacement of oxygen by OH^- or F^-. The Si for Al replacement is the one that takes place in the solid solution of $Na_2Si_3O_7$ in gehlenite but cannot do so in akermanite. Phosphorous analyses have not been made on natural melilites; but, if phosphorous is present, it is not at all likely that it exists in appreciable amounts; therefore, it cannot be responsible for the presence of any significant amount of soda in the structure. The third possible anion replacement, that of Al for Mg, does not take place, as is shown by the results obtained with the mix of the composition Ak 45 per cent–Geh 45 per cent–$Na_2Si_3O_7$ 10 per cent. The Al for Mg replacement in akermanite would result in the formula $NaCaAlSi_2O_7$, which is nothing more than a composite of gehlenite and $Na_2Si_3O_7$. Inasmuch as the three-"component" mix, described above, contains this "molecule" as well as akermanite and yet shows no tendency to form a solid solution over and above that formed in the gehlenite, it is obvious that Al does not replace Mg in akermanite.

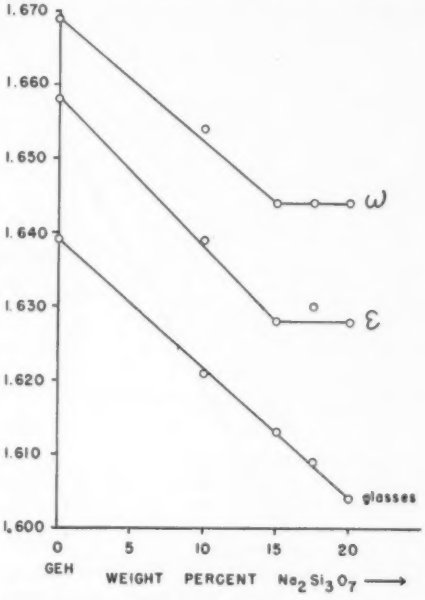

FIG. 2.—Refractive indices of melilites and glasses.

The importance of OH^- in replacing $O^=$ with a resulting Na^+ for Ca^{++} replacement in melilites cannot be evaluated with the analytical data on hand. It has been observed that in Si_2O_7 structures OH^- does not replace $O^=$. This is an empirical fact and has no theoretical basis. On this evidence alone one might say that melilites are not likely to show this substitution. Some of the published analyses show an appreciable water content (in one case up to 1.59 per cent), but

in most cases H_2O was not determined. Much of the water found in ordinary analyses probably is not present as the hydroxyl; some may come from impurities and some may be adsorbed. The ordinary means of determining H_2O is not satisfactory, as it does not detect all the OH^- that may exist in the crystal; it is necessary to break down the structure completely (for example, see Sahama, 1946) before all the hydroxyl is liberated as water. One per cent of water, present as OH^- replacing oxygen, can account, however, for approximately 3 per cent of soda, and water, therefore, should be carefully determined in the course of any analytical work that is concerned with crystal chemistry.

With the possible exception of this cation replacement, it is thus improbable that soda can enter the melilites by any means other than as a replacement for Ca with the accompanying Si^{4+} for Al^{3+} substitution. This is tantamount to saying that the $Na_2Si_3O_7$ "molecule" is the only sodic "molecule" that can act as a substitutive end-member. The fact that the experimental results show that only 3.85 per cent Na_2O can be taken up by pure gehlenite and that the soda content drops as the akermanite content of the crystals increases makes it difficult to account for the high values of Na_2O reported in natural melilites if water is not considered. It must also be remembered that, at the rather high temperatures reached in the laboratory investigation, it is likely that more solid solution takes place than is probable in nature.

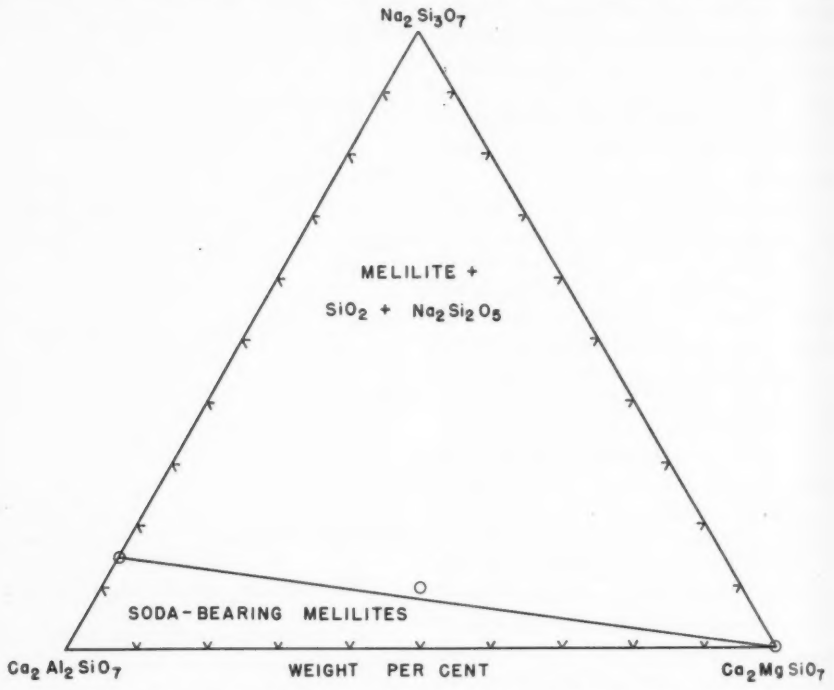

FIG. 3.—The stability field of the soda-bearing synthetic melilites

The problem of the soda content of

natural melilites becomes more confusing when the chemical analyses are studied. Berman (1929) states that "the gehlenites in general have a low percentage of the minor molecules $Na_2Si_3O_7$ and $CaSi_3O_7$." Figure 4 is a plot of soda (+ potash) against magnesia (+ ferrous iron). The abscissa represents the gehlenite-akermanite ratio in simple (synthetic) melilites, as pure gehlenite contains no MgO and pure akermanite contains 14.79 per cent by weight. The previously discussed limit of soda in this system is represented by the line from 3.85 per cent at zero MgO to zero per cent at 14.79 per cent MgO. Twenty-three analyses collected by Berman (1929) are plotted, plus the Na_2O–MgO ratios of two analyses done since 1929 (Tilley, 1929; Schaller, 1942). Six melilites have less than ½ per cent soda, and five of these are rich in the gehlenite "molecule." The sixth is virtually pure akermanite and would be expected to show little or no Na_2O. Indeed, there is shown a weak tendency for the more akermanitic melilites to be the richest in Na_2O, which is just the opposite of what would be expected on the basis of the experimental and theoretical considerations. Figure 4 clearly illustrates the soda problem in the melilites. The Na_2O that can be accounted for as $Na_2Si_3O_7$ varies from 3.85 per cent in gehlenite to zero per cent in akermanite; seventeen of the twenty-five points fall above the line delimiting this soda content.

Berman (1929) pointed out the fact that the Ca + Na content is generally below the value of 20 expected on the basis of 70 oxygen atoms. It was because of this deficiency that he proposed $CaSi_3O_7$ as an end-member. Inasmuch as Na must replace Ca and because $CaSi_3O_7$ (or any calcium silicate for that matter) does not form a solid solution with the melilites, the value of 20 atoms of Ca + Na would be expected to be constant. Figure 5 is a plot of Ca against Na (+ K). The units are atomic composition, on the basis of 70 oxygen atoms in the melilite formula. Replacement of Ca by Na is on a 1:1 basis, so that perfect substitution would result in the straight line with the negative slope of 1 shown on the diagram. The 23 analyses considered by Berman are here plotted; the tendency for the Na + Ca values to be less than 20

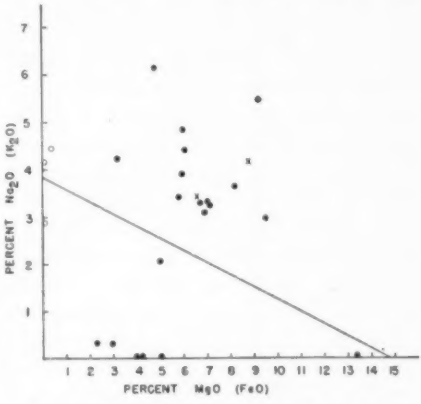

FIG. 4.—Na_2O(+ K_2O) versus MgO(+ FeO), in weight per cent, in natural melilites. Crosses are analyses since 1929. Open circles are called "sarcolites."

is obvious—only 4 points lie above the line and 4 lie on it.

The fundamental problems of the melilites have not been solved by this investigation, but it is hoped that some of the points have been clarified. The deficiency in total Ca + Na cannot be readily explained; the soda in excess of that shown to be taken up by gehlenite cannot be explained here. Berman showed that the Y_3 member of his melilite general formula (i.e., Si, Mg, Al) is quite constant, varying but slightly from the theoretical value of 30 atoms per 70 oxygen atoms. Poor chemical analyses (most of these are quite old) might be blamed for the discrepancies except for this con-

stancy of Mg, Si, and Al. The same can be said for inclusions or impurities in the samples. The influence of pressure is unknown, but its effect must be small, as none of the melilites examined were formed under conditions of very high pressure. The influence of hydroyxl is unknown but probably not large, as al-

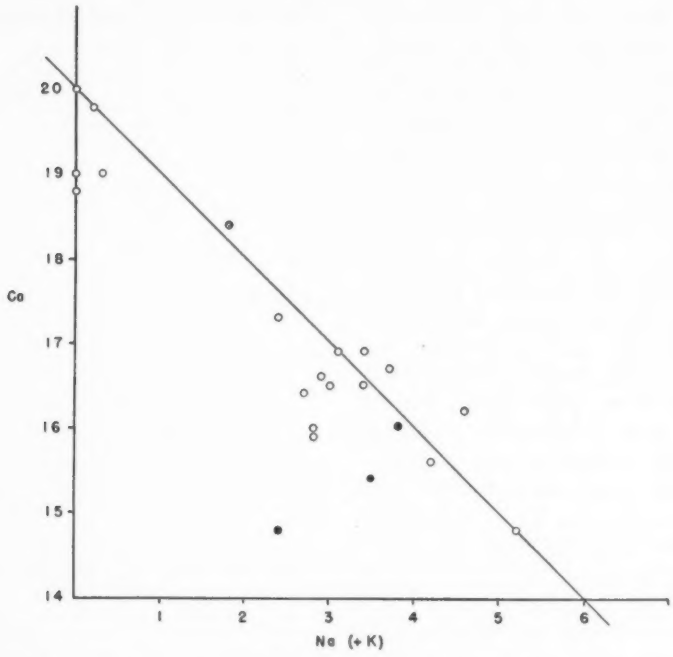

FIG. 5.—Atomic ratios of Ca versus Na(+ K) in natural melilites. Solid circles represent "sarcolites"

ready discussed. It should also be noted that if OH^- substitutes for $O^=$, the soda substitution then occurs with no effect on the Si content. The fact that the melilites tend to show high Si contents might be used as an argument against the hydroxyl substitution and might favor the argument for $Na_2Si_3O_7$ solid solution.

More information is necessary on the natural minerals; careful chemical and X-ray work should be done. An X-ray investigation of the synthetic series may also prove fruitful. The possibility exists that there may be significant differences between the metamorphic and the igneous melilites, a problem that offers possibilities of profitable study with relation to other minerals as well.

THE POSSIBILITY OF A CALCIUM-ALUMINATE SOLID SOLUTION IN GEHLENITE

In a recent paper dealing with a portion of the Na_2O–CaO–Al_2O_3–SiO_2 system (Goldsmith, 1947) the possibility of a solid solution of $CaO \cdot 2Al_2O_3$ in gehlenite was considered. The improbability of obtaining mix-crystals of gehlenite and the aluminate was pointed

out, as a substitution of the small Al ion for the large Ca ion is involved.

Ca Ca $AlAlSiO_7$ (gehlenite)
Ca Al $AlAlAlO_7$ ($CaO \cdot 2Al_2O_3$)

The substitution of Al for Ca, as far as the writer knows, is not recognized in any known structure.

It was felt that the problem was interesting enough to warrant an investigation, and, accordingly, glasses were prepared in the gehlenite-rich portion of the system $Ca_2Al_2SiO_7$–$CaO \cdot 2Al_2O_3$. The glasses were devitrified at approximately 1,300° C. and examined microscopically. Mixtures up to 30 per cent $CaO \cdot 2Al_2O_3$ were optically homogeneous, i.e., but a single phase could be detected. The material devitrified as a somewhat fibrous, finely crystalline product with parallel and subparallel crystallographic orientation, as is common when a glass is crystallized well below its solidus temperature. A rather small lowering of the refractive indices was noted; the 30 per cent $CaO \cdot 2Al_2O_3$ mix showed $N = 1.658$, $n = 1.645$, a drop of approximately 0.01 from the indices of pure gehlenite.

No solid solution between gehlenite and any calcium aluminate was found in the original work on the system CaO–Al_2O_3–SiO_2 (G. A. Rankin and F. E. Wright, 1915). The eutectic between gehlenite and $CaO \cdot 2Al_2O_3$[2] is located at 31 per cent $CaO \cdot 2Al_2O_3$, and it appeared to the writer that even if the gehlenite–$CaO \cdot 2Al_2O_3$ solid solution existed, it is improbable that it would extend all the way to the eutectic. In the earlier stages of this work, on compositions low in $CaO \cdot 2Al_2O_3$, the apparent optical homogeneity led to the conclusion that solid solution between gehlenite and the calcium aluminate certainly took place. It was on this basis that an abstract was published,[3] but additional work has cast considerable doubt on the validity of the above conclusion.

The suspicion that the optical data were perhaps misleading led to an investigation of the devitrified products by means of X-ray diffraction. Powder pictures were made of pure gehlenite, pure $CaO \cdot 2Al_2O_3$, and of a mechanical mixture of 90 per cent gehlenite and 10 per cent $CaO \cdot 2Al_2O_3$. It was found that considerably less than 10 per cent $CaO \cdot 2Al_2O_3$ can be detected in gehlenite in X-ray powder pictures, because the three strongest calcium-aluminate lines were easily detected in the picture of the composite preparation. Powder pictures were then made of the mixtures that had been devitrified at approximately 1,300° C., with the result that the 90 per cent gehlenite–10 per cent $CaO \cdot 2Al_2O_3$ showed none of the lines of the $CaO \cdot 2Al_2O_3$, whereas those of 20 per cent and higher $CaO \cdot 2Al_2O_3$ content did. These data revealed that microscopic examination was insufficient in this system. A submicroscopic intergrowth of gehlenite and $CaO \cdot 2Al_2O_3$ apparently formed when the glasses were devitrified at 1,300° C., which appeared to be optically homogeneous. It also appeared, however, that some solid solution was indeed taking place, as the 10 per cent mixture did not show calcium-aluminate lines. If any line shifts in the powder pictures are present, they are so small as to be within the limits of error of measurement.

The liquidus relations (originally determined with $3CaO \cdot 5Al_2O_3$ as one end-member by Rankin and Wright [1915]) were not reinvestigated because of the high temperatures involved. Because the X-ray evidence on the 10 per

[2] Erroneously identified as $3CaO \cdot 5Al_2O_3$ in the original investigation and subsequently labeled as such in all later publications and diagrams in this country.

[3] Geol. Soc. America Bull. 58, no. 12 (pt. 2), December, 1947, p. 1183.

cent $CaO \cdot 2Al_2O_3$ mix indicated some solid solution, it was thought worth while to determine the solidus curve. In the attempt to locate the solidus, using both previously devitrified material and glass, all mixtures showed marked inhomogeneity when held at high temperatures. The 10 per cent mix, for example, which was homogeneous at 1,300° C., showed numerous tiny blebs and inhomogeneities at 1,500° C., and the X-ray picture indicated the presence of $CaO \cdot 2Al_2O_3$. The phenomenon of homogeneity and apparent solid solution at 1,300° C., with inhomogeneity at higher temperatures, is indeed unusual, the opposite of what normally would be expected. The X-ray evidence of the presence of $CaO \cdot 2Al_2O_3$ in mixtures with more than 10 per cent of the component $CaO \cdot 2Al_2O_3$ cannot thus be explained by unmixing at low temperatures from a high-temperature solid solution, as it is obvious that little or no solid solution exists at near-solidus temperatures. A solid solution stable at low temperature and unstable at higher temperatures is difficult to explain unless one assumes an inversion in gehlenite itself; no evidence for such an inversion has ever been noted.

Intimate mixtures of very finely ground crystalline gehlenite and $CaO \cdot 2Al_2O_3$ were heated in an attempt to produce reaction in the solid state. If solid solutions were formed between these components, it might be expected that the end-members would react in the solid state to produce mix-crystals. Heating for 3 days at various temperatures up to 1,500° C. produced no change in microscopic appearance or X-ray diffraction pattern; the two phases could easily be detected (90 per cent gehlenite, 10 per cent $CaO \cdot 2Al_2O_3$). However, these results do not conclusively prove that solid solution cannot take place. It is not improbable that a solid-state reaction of this sort might not always take place (or do so very slowly), particularly if the solid solution is a rather unstable one. It is not unlikely that mix-crystals might be obtained by direct crystallization from a melt (glass) in which but a very small rearrangement of atomic positions is necessary; but if the solid solution is a relatively unstable one, the necessary amount of rearrangement might not take place from the two stable crystalline phases. To do so would require the destruction of the lattices of the two crystalline phases, with relatively large-scale migration and construction of a new single phase.

Experimental work performed on the subsolidus relations between $Ca_2Al_2SiO_7$ and $CaO \cdot 2Al_2O_3$ thus shows rather conflicting evidence on the question of solid solution. Devitrification of glasses at approximately 1,300° C. apparently results in a single phase, whereas at 1,500° C. an obviously inhomogeneous product results. It appears that at 1,300° C. a solid solution with an upper limit of approximately 10 per cent $CaO \cdot 2Al_2O_3$ may form, although several other observations tend to contradict this conclusion. No line shifts or intensity changes could be detected in the X-ray powder pictures, no observable reaction in the solid state takes place in 3 days' time, and little or no faith can be put in optical examination of this system.

If there is solid solution of a calcium aluminate in gehlenite, it is almost certainly not present in the rock-forming minerals. The melilites are not undersilicated but generally tend to show excess silica, as mentioned earlier. It has also been stated that the Y term of Berman (Si + Mg + Al) is constant at a value of 30, and such would not be the case if $CaO \cdot 2Al_2O_3$ were present in the crystals.

The intergrowths of $CaO \cdot 2Al_2O_3$ in gehlenite that could not be resolved opti-

cally are interesting if one considers that such intergrowths may be present but unrecognized in certain natural minerals. The possibility of fine intergrowths giving misleading information on chemically analyzed minerals should be considered.

SUMMARY AND CONCLUSIONS

The solid solution of $Na_2Si_3O_7$ in gehlenite ($Ca_2Al_2SiO_7$) was experimentally determined and takes place to the extent of 15 per cent by weight (3.85 per cent as Na_2O). Soda cannot enter the structure of pure akermanite, and the amount of Na_2O that can enter mixcrystals of synthetic gehlenite and akermanite varies in a linear fashion from 3.85 per cent in gehlenite to 0 in akermanite. The high soda content found in some natural melilites (up to approximately 6 per cent) cannot be accounted for, on the basis of this investigation, by solid solution of $Na_2Si_3O_7$ in the melilites. It is shown that the only reasonable way that soda can enter the melilite structure, other than as the "end-member" $Na_2Si_3O_7$, is by OH^- replacement of $O^=$; the magnitude of this replacement has not been fully evaluated. The phase diagrams of gehlenite-wollastonite and akermanite-wollastonite show no solid solution complications and are truly binary in nature, indicating that the melilites do not take any calcium silicate (of any formulation) in solid solution. This fact, plus structural considerations, eliminates any of the earlier proposed lime endmembers, such as Berman's "submelilite," $CaSi_3O_7$. Additional careful chemical (including OH^- and F^- determinations) and X-ray work on natural melilites is necessary.

The possibility of the solid solution of $CaO \cdot 2Al_2O_3$ in gehlenite is considered. There is some indication that approximately 10 per cent $CaO \cdot 2Al_2O_3$ may be taken up by gehlenite, but the experimental evidence is somewhat contradictory. If such a solid solution does exist, it is indeed unusual in that the substitution of Al for Ca is involved.

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RADIAL DIFFUSION AND CHEMICAL STABILITY IN THE GRAVITATIONAL FIELD[1]

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ABSTRACT

Because gravity influences chemical equilibrium, consideration of the thermodynamics of our globe requires understanding of chemical processes and stability in the gravitational field. The influence of gravity on stability of mixed condensates (crystals or magmas) and of stoichiometric crystals is analyzed. The theoretical analysis supports the hypothesis of granitization and metasomatism by upward diffusion of some elements.

INTRODUCTION

In recent years an increasing number of students of granitization have realized that these phenomena cannot be completely accounted for without assuming large-scale diffusion through the solid rocks (Wegmann, 1935; Roubault and Perrin, 1939; Backlund, 1946; Barth, 1948*b;* Ramberg, 1944*a*, *b*, 1945).

In the earth the principal laws governing radial diffusion have never, so far as I know, been fully understood by geologists. Some years ago the writer (1944*a*, 1945, 1946) proved that mechanical instability in the gravitational field results in chemical or thermodynamic instability. In other words, if rocks or magmas with comparatively high density rest on rocks or magmas with low density, then chemical potential differences exist, tending to make the two bodies change places with the help of dispersion, vertical diffusion, and consolidation.

This paper is devoted to the principal laws of radial diffusion in the earth. We must base our estimate of the role of diffusion in the metasomatic alteration of the earth's crust in part on these principles.

In the earth's crust and even deep in the earth, most material is condensed and contains only negligible quantities of free gases. Therefore, chemical stability in the earth can be said to be equivalent to equilibria of diffusion. Such equilibria are not characterized by diffusion of matter in the form of activated molecular particles (atoms, ions) in but one direction, even if infinite intervals of time are considered. The equilibrium is believed to be dynamic, that is, equal numbers of activated molecular particles migrate in opposite directions through any given cross section in the earth.

Like other chemical processes, the diffusion currents are driven by chemical forces which can be expressed in terms of the chemical potential, μ; free energy, ζ; activity, a; fugacity, f; and vapor tension, π. These terms are not identical but are related by means of more or less simple equations.

As pointed out in previous papers (1944*a*, 1946), the writer finds it most convenient to make use of the term "partial vapor tension," π, if both diffusion and chemical equilibrium are involved and even if the system is completely condensed. The term "partial vapor tension" is accordingly considered an important thermodynamic property of condensates,

[1] Presented at the University of Oslo, April, 1946, as the third part of a series on the "Thermodynamics of the Earth's Crust." The first two parts have already been published (Ramberg, 1944*b* and 1945). Manuscript received December 15, 1947.

being useful whether a vapor is formed or not.

Because some geologists (I. Oftedal, 1947) have criticized the concept of vapor tension of minerals as indicative of the chemical potential, the views of the distinguished physicochemists, Lewis and Randall, are significant (Marsh, 1935, p. 31):

> One of the most striking results of this character is obtained if we calculate the vapor pressure of tungsten at 25° from experiments at very high temperatures. The result, 10^{-149} atmospheres, would mean that the concentration of tungsten vapor would be less than one molecule in a space equivalent to the known sidereal universe. Such a calculation need not alarm us. Allowing for the possibilities of experimental uncertainties, we may use such a calculated vapor pressure in our thermodynamic work with the same sense of security as we use the vapor pressure of water.

Even in crystalline systems, diffusion resistance is never infinite (Barrer, 1941), e.g., horizontal chemical potential or vapor-tension gradients will always give rise to diffusion of activated particles, although the rate of this process in some cases will be very slow (Jost, 1937). Chemical equilibrium in a horizontal plane in the earth's crust is therefore, as we know, characterized by no gradient of the partial chemical potentials.

Equations of the following type:

$$\frac{dm}{dt} = -\,q\,D'\,\frac{d\pi}{dx} \qquad (1)$$

hold for the diffusional transfer and along a horizontal plane in the earth. The quantity dm/dt is the power of diffusion through section q; D' is a diffusion coefficient depending on the character of the system, on P and T, and on the type of migrating particles. In heterogeneous systems such as rocks, D' is to be considered a mean value depending on the diffusion coefficients for the intergranular diffusion, the mosaic-fissure diffusion, and the volume diffusion in the different phases (minerals, pore solutions). The quantity $d\pi/dx$ is the partial tension gradient in the horizontal direction.

Obviously, stability exists when $dm/dt = 0$, i.e., when $d\pi/dx = 0$, since q and D' are finite quantities.

Considering, now, the effect of gravity, it is clear that, in order to insure chemical stability in a vertical direction, the gravitational attraction affecting the diffusing activated particles must be compensated by a vapor-tension gradient acting in opposition to gravitation.

Chemical stability in the gravitational field is achieved if the partial vapor tension of a given element in the crystalline mantle of the earth increases with depth in accordance with the simple equation (Ramberg, 1944*a*),

$$\pi_y = \pi_0\, e^{(My)/(RT)}, \qquad (2)$$

where π_y is the partial vapor tension at depth y; π_0 is the stable partial tension at level 0 (e.g., the surface); M is molecular weight of the particle considered; and R is the gas constant. Hence, the stable partial tension gradient is

$$\frac{d\pi}{dy} = \pi\,\frac{M}{RT}, \qquad (3)$$

when y is positive downward.

Partial tension gradients steeper than those corresponding to equation (3) make the corresponding elements diffuse upward through the crust because in this case the upwardly directed chemical force is stronger than the gravitational attraction. On the other hand, diffusion has to go downward if the partial tension gradient is less than that expressed in equation (3).

Diffusion parallel to the gravitative force-lines can accordingly be expressed by an equation of the following form:

$$\frac{dm}{dt} = -\,q\,D'\left(\frac{d\pi}{dy} - \pi\frac{M}{RT}\right). \qquad (4)$$

This equation demonstrates that dm/dt is zero (i.e., chemical equilibrium exists) when $d\pi/dy = \pi(M/RT)$, and that the diffusion current is running upward if $d\pi/dy > \pi(M/RT)$, and downward if $d\pi/dy < \pi(M/RT)$.

Considering the chemical potential μ, equilibrium conditions are characterized by no change in μ throughout the system (MacInnes, 1939, p. 174). This means that the relation between vapor tension and chemical potential changes with position in a gravitational field.

THE STABILITY OF MIXED CRYSTAL AND MAGMA LAYERS IN THE GRAVITATIONAL FIELD

Let us first consider a thick layer of an ideal mixed crystal, say A_mB_n, in the gravitational field. For convenience, A and B are elements, e.g., metals. Furthermore, let the temperature be constant and the composition of the top of the layer (which does not necessarily reach up to the earth's surface) be A_mB_n.

According to Rault's law, the partial vapor tension of the ideal mixed crystal at the top will be

$$\pi_{A_m} = \pi_A \frac{m}{n+m} \tag{5}$$

for the partial A-tension, and

$$\pi_{B_n} = \pi_B \frac{n}{n+m} \tag{6}$$

for the partial B-tension. Here π_A and π_B are the vapor tensions of the metals A and B in their pure state; m is the number of mols of A in the mixed crystal; and n the number of mols of B. If there is stability throughout the whole layer, the partial tensions at every depth, y, must coincide with the law of stability (eq. [2]) viz:

$$\pi_{A_y} = \pi_A \frac{m}{n+m} e^{M_A y/RT} \tag{7}$$

and

$$\pi_{B_y} = \pi_B \frac{n}{n+m} e^{M_B y/RT}, \tag{8}$$

where π_{A_y} and π_{B_y} are the stable partial tensions with reference to A and B in the depth y; and M_A and M_B are the atomic weights of A and B, respectively.

The pressure gradient will make the partial tension increase downward, but only in the case of monatomic condensates will this downwardly increasing vapor tension harmonize with the law of chemical stability in the gravitational field (eq. [2]). In mixed crystals there will be a continuous change in composition.

The mechanical pressure at depth y is $P_y = d \cdot y$ where, for convenience, the density d is supposed to be independent of both pressure and change in composition of the mixed crystal layer.

According to the equation expressing the relation between vapor tension and mechanical pressure, the pressure in depth y makes the partial vapor tension rise in the following manner, provided that composition is constant:

$$\pi_{A_P} = \pi_A \frac{m}{n+m} e^{P_y V_A/RT} = \pi_A \frac{m}{n+m} e^{d \cdot y V_A/RT} \tag{9}$$

and

$$\pi_{B_P} = \pi_B \frac{n}{n+m} e^{P_y V_B/RT} = \pi_B \frac{n}{n+m} e^{d \cdot y V_B/RT}, \tag{10}$$

respectively. Here V_A and V_B are the fictive molal volumes of the components A and B, respectively, in the mixed crystal. (The "fictive molal volume" of a given component in a mixture is the change in volume that an infinitely large body of the mixture undergoes by reversible loss or gain of 1 mol of the considered component.) The fictive molal

volume of a certain ion or atom is different in different phases. Likewise, it changes with changing composition and physical condition of a given phase.

It follows that if equation (7) is identical with equation (9) and equation (8) is identical with equation (10), then the stable conditions correspond to a homogeneous mixed crystal layer. On the other hand, if these equalities do not exist, i.e., if,

$$\pi_{A_P} \gtrless \pi_{A_y}, \quad \text{or} \quad d \gtrless \frac{M_A}{V_A}, \quad (11)$$

and

$$\pi_{B_P} \gtrless \pi_{B_P}, \quad \text{or} \quad d \lesseqgtr \frac{M_B}{V_B}, \quad (12)$$

then constant composition of the layer will result in a chemically unstable state.

If $d > M_A/V_A$ and $d < M_B/V_B$, there will exist forces which make the element A migrate upward and B downward. If $d < M_A/V_A$ and $d > M_B/V_B$, then A migrates down and B up. The change in composition at different levels thus engendered leads to the following conditions: $\pi_{A_P} = \pi_{A_y}$ and $\pi_{B_P} = \pi_{B_y}$, obviously due to the alteration of the term $n/(n+m)$ and $m/(n+m)$ in equations (9) and (10). Thus we find the stable composition, $(m/[n+m])_y$, at depth y when the surface composition is $m/[n+m]_0$

$$\left(\frac{m}{n+m}\right)_y = \left(\frac{m}{n+m}\right)_0 e^{(M_A y - d \cdot y V_A)/RT} \quad (13)$$

(see also MacInnes, 1939, p. 175). The term M/V thus governs the change in composition with depth in ideal mixed crystals. I have named this quantity the "fictive density" of the particle (ion, atom) in question (Ramberg, 1946). In harmony with what is said above about the fictive molal volume, the fictive density of a certain element or ion changes from one crystal to another. If the fictive densities of elements in a substitution mixed crystal of type A_mB_n differ from the density of the crystal itself, then the forces of differentiation will try to change the composition in a vertical direction in the gravitational field. Elements with fictive densities larger than the density of the crystal will be gradually concentrated at lower levels and vice versa.

In silicate minerals or melts these principles may be applied to elements that mutually replace each other. Because such elements have nearly equal fictive molal volume, their fictive densities are approximately proportional to the atomic weight. Consequently, in a number of ferromagnesian minerals and in mixed crystals of the plagioclase types, the stable condition is characterized by a gradual downward increase in the Fe/Mg ratio and the Ca/Na ratio, respectively. Provided that the fictive molal volumes of the mutually replacing elements are equal, equation (13) leads into:

$$\left(\frac{m}{n}\right)_y = \left(\frac{m}{n}\right)_0 e^{[(M_A - M_B)y]/RT}. \quad (14)$$

This equation can be used with fair accuracy for calculating the change in the Ca/Na ratio with depth in the plagioclase mixed crystal series. Assume, for example, a temperature of 400° K. and Ca/Na = 1 at the surface, there is, at a depth $y = 20$ km., a stable ratio of Ca/Na = 2.7. In ferromagnesian minerals a change of Fe/Mg ratio from 1 to 2.7 extends from the top to about 9 km. depth.

In addition-mixed crystals, some elements may enter vacant places in the lattice and give rise to only a small expansion of the unit cell, i.e., the fictive molal volume may in this case be very small (in very rare cases it may be zero

or even negative). This means that the fictive density of the elements that occupy the vacant places may be very large, even if the elements have a small atomic weight. At stable conditions the incorporated elements in this case will be concentrated at lower levels, in spite of being incorporated in a dense crystal.

In mixed crystals of metal and hydrogen the unit cell does not necessarily increase through the absorption of hydrogen because the small H-atoms find their place in the interstices among the metal atoms. In some cases even, contraction may take place, resulting in a negative fictive molal volume of hydrogen so that the H_2-tension of the hydrogen-metal mixture decreases with increasing external pressure (eq. [9]). Therefore, if hydrogen-metal mixtures exist in the earth, the very light hydrogen atoms will be concentrated at lower levels. The degree of concentration, however, is obviously limited by the miscibility between metal and hydrogen.

These peculiarities must be taken into consideration when speculations as to the condition of our globe's deeper parts are discussed (Kuhn and Rittmann, 1941).

THE GRAVITATIVE STABILITY OF STOICHIOMETRIC CRYSTALS

In this type of crystal there are very narrow fields of homogeneity, so that the element ratio cannot diverge much from the stoichiometric ratio. On the other hand, even very negligible change in composition gives rise to considerable change in partial vapor tension. In a system of elements which are able to combine into different stoichiometric solid compounds, gravitation will obviously tend to develop a mechanically stable arrangement of different layers with different compositions and different densities. This process will take place with the help of bodily up-and-down motion of rocks and magmas in the earth.

It has, however, never been fully recognized by geologists that such a mechanically stable arrangement in the earth also corresponds to thermodynamic stability. This means that any large-scale structure of the earth not corresponding to a stable concentric-layer arrangement possesses potential chemical forces tending to develop the stable arrangement. In this connection it cannot be overemphasized that the only process able to equalize chemical instabilities in the earth is diffusion.

To explain the principal features, let us consider a theoretical system consisting of the elements A and B which can combine into a stoichiometric phase $C = AB$. The atomic weight of A is supposed to be less than that of B: $M_A < M_B$. The densities of A, B, and C are d_A, d_B, and d_C, respectively, where $d_A < d_C < d_B$.

The vapor tension of A is π_A and of B, π_B at $P = 0$ and a given temperature. C is supposed to have a partial A-tension equal to π_{C_A} and a partial B-tension equal to π_{C_B} at $P = 0$. Now $\pi_{C_A} < \pi_A$, and $\pi_{C_B} < \pi_B$, so that the reaction:

$$\text{High } T,\ T\mu,\ \text{Low } T$$
$$A + B \rightarrow C\,, \qquad (14)$$

goes to the right at zero pressure and at the temperature under consideration.

As in mixed crystals, there will exist an instability in a homogeneous thick C layer in the earth if

$$\frac{M_A}{d_C} \gtrless V_{AC} \quad \text{and} \quad \frac{M_B}{d_C} \lessgtr V_{BC},$$

where V_{Ac} and V_{Bc} are the fictive molal volumes of A and B in the compound C. The element with fictive molal volume larger than M/d_C tends to migrate upward and vice versa.

In this example, $V_{Ac} > M_A/d_C$ and $V_{Bc} < M_B/d_C$, so that A will tend to migrate toward the top and B downward. These migrations in combination with phase transitions will split up the homogeneous C layer, creating a stable arrangement of three layers with A on the top, C in the middle, and B at the bottom. The thicknesses of the layers in this stable arrangement depend on the character of the system, on temperature, and on the gravitational force. At temperatures just below the transition temperature of equation (14) (at $P_2 = 0$) the stable C layer will be thin, but at lower temperatures the thickness of C may be very large at stable conditions.

It is also of interest to consider the case of the compound $C = (AB)$ with a density less than d_A due to a loose-packed lattice of C: $(d_C < d_A < d_B)$. In this case the potential mechanical energy is at a minimum, and mechanical stability exists when C exists as a top layer above A and B lies at the bottom. It can also be proved that this arrangement represents the true chemical stability, otherwise a circular process arises (Ramberg, 1944*a*).

If such a light C layer is exposed to the differentiational effect of gravity, then dispersion, diffusion, and consolidation will develop an A layer in the middle parts of the original C layer, while B atoms migrate farther down, consolidating as a pure B layer at the bottom.

In our first example $(d_A < d_C < d_B)$, we may have originally a thick A layer lying above a B layer. Then reaction and diffusion across the boundary generate an intermediate C layer, which eventually attains the stable thickness. The same original arrangement in case 2 (in which $d_C < d_A < d_B$) will not give rise to interaction of A and B along the boundary (provided that the A layer is thick enough to create a sufficiently high pressure along the boundary), but B migrates up through the A layer, unable to form C in the lower parts of the A layer. Above a certain level in the A layer, however, C begins to generate until a stable C layer is formed at the top. Similarly, if $d_C > d_B > d_A$, we find that diffusion processes tend to rearrange the system, leaving C at the bottom and A and B above.

The system Fe–O_2 and their compounds, with which the writer has briefly dealt previously (Ramberg, 1946) and which Barth has tried to treat quantitatively in this journal (1948*a*), harmonizes with the above considerations.

In a sufficiently thick layer of magnetite in the earth, there exist forces tending to make oxygen migrate upward and iron downward if the fictive molal volumes of iron and oxygen satisfy the following requirement:

$$V_{O_{mt}} > \frac{M_O}{M_{Fe}} V_{Fe_{mt}}, \qquad (15)$$

where $V_{O_{mt}}$ and $V_{Fe_{mt}}$ are the fictive molal volumes of O and Fe in magnetite.

Although the fictive molal volumes may differ considerably from the so-called "ionic" or "atomic" volumes of the different elements, it seems very reasonable that, for magnetite, equation (15) is satisfied and also that $V_{O_{mt}} > V_{Fe_{mt}}$. Thus gravity tends to differentiate the magnetite layer, establishing an arrangement of O_2 gas on the top (the pressure of this gas corresponds to the oxidizing pressure of magnetite at the given temperature), hematite layer below, magnetite layer in the middle, FeO farther down, and iron in its pure state at the bottom.

Because magnetite actually has somewhat lower density than hematite, however, there exists a small instability in this arrangement. The true chemical and

mechanical stability exists when the magnetite and hematite layers exchange places, so that the less oxidized and less dense phase is in contact with the O_2 atmosphere. This may seem impossible at first glance, but the O_2 tension at the surface in the first-mentioned arrangement equals the oxidizing tension of the reaction $2Fe_3O_4 + \frac{1}{2}O_2 \rightleftarrows 3Fe_3O_3$ at the given T. Because the rearrangement from the unstable to the true stable state is combined with a certain depression of the O_2 tension throughout the whole system, it is clear that magnetite will not be oxidized along the boundary against the dilute O_2 gas. The O_2 tension above the magnetite layer is somewhat lower than the oxidizing tension, whereas the oxygen tension below a certain depth is, at stable conditions, able to oxidize the magnetite into a hematite phase.

The processes in the original homogeneous magnetite layer may be explained as follows: Fe will diffuse downward from the middle parts of the layer considerably more rapidly than O migrates upward, so that the iron-poor phase, hematite, is formed in the middle parts of the original layer. Oxygen is comparatively strongly bound to the hematite phase, which absorbs the O particles migrating up from the deeper levels; only a few oxygen ions or atoms are able to escape from the hematite layer and eventually reach the surface.

Whatever the pure kinetics of the process, it is clear that the ideal stable arrangement of a Fe-O_2 system in the gravitational field is not a homogeneous iron oxide phase but several layers arranged as explained above. However, a minimum thickness of the original homogeneous magnetite or hematite is required to start and maintain the differentiation process. Likewise, the thicknesses of the several established layers are also definite values depending on composition of the system (the mean Fe/O ratio) and on temperature.

With certain modifications, this ideal system is of interest in discussion of the degree of oxidation in the earth at different levels. Temperature, pressure, and chemical environments are not the only factors influencing the degree of oxidation (Goldschmidt, 1943); the differentiational effect of gravity plays an important role, even in crystalline and condensed systems.

This idea of chemical squeezing of some elements with low fictive density out of the deeper parts of the earth was presented some years ago (Ramberg, 1944*a*, 1946). The degree of oxidation at different levels in the crust or the ratio of oxygen to metallic ions was considered partly due to upward diffusion of oxygen and downward migration of elements with high fictive density. Later Barth (1948) calculated the ratio between oxygen and metallic atoms in rocks from different depths and arrived at a conclusion in full agreement with the present writer. It is also interesting that Harrison Brown and Claire Patterson recently (1948) arrived at similar ideas concerning the content of oxygen in meteorites and in the earth. They concluded from investigations of meteorites that oxygen might be squeezed out from deeper parts of the planets.

In the earth today the atmosphere O_2 tension is very much higher than the oxidizing tension of magnetite at low temperatures. It is therefore impossible to assume that O_2 will migrate out of the earth's surface owing to gravitative differentiation in a solid earth at the present comparatively low temperatures of the crust. Provided that temperature was sufficiently high in an earlier evolutionary stage of the earth, gravitative differ-

entiation might have contributed to the formation of an oxygen atmosphere by chemical squeezing of oxygen out of deeper levels in a molten or solid earth. In more recent times similar processes will take place at greater depths, where temperature is high and the effect of the atmospheric oxygen does not disturb the processes.

Other examples of the influence of gravitation on chemical stability are the hydrous minerals, such as micas, hornblende, and others. Gravitation makes such minerals chemically unstable below certain depths, owing to the chemical squeezing of H_2O out of the lattices and the upward migration of the water or independent H^+ and O^{--} ions.

If hydrous minerals, e.g., hornblende, biotite, or muscovite, exist below a certain level (which depends on temperature), then the superincumbent load raises the partial H_2O tension (owing to the large fictive molal volume of OH^- in the lattice) above that which corresponds to vertical diffusional stability. Consequently, forces exist that make H_2O (or $H^+ + O^{--}$) migrate toward the top, eventually resulting in hydrothermal metasomatism. Below a certain depth, which seems to be impossible to calculate because of lack of data, the chemical squeezing of water is so intense that hydrous minerals like pyroxene, garnets, etc., form at the cost of mica and hornblende.

In micas a chemical squeezing of the large potassium ions will take place; the potassium tends to migrate upward, to a small extent contributing to the granitization process.

This process is of interest for the formation of the *granulite facies* of rocks. The chemical squeezing-out of the water of the hydrous minerals at high pressure and great depths indicates that high temperature is not necessary for the formation of the anhydrous minerals of the granulite facies.

The existence of olivine-rich rocks at great depths may be explained to some extent in a similar way. At a certain depth, Si and O_2 will be squeezed out of the hypersthene lattice. Upward migration of these elements will take place, resulting in Si-metasomatism at higher levels and olivine formation at lower levels.

Finally, let us consider the differential effect of gravitation on the well-known reaction:

$$\text{Calcite} + \text{quartz} \rightleftarrows \text{wollastonite} + CO_2 . \qquad (16)$$

At the low CO_2 tension existing in the air, wollastonite is formed at about 500°. In closed systems where CO_2 is exposed to the total rock pressure the reaction temperature rises with increasing pressure, according to V. M. Goldschmidt (1912).

The wollastonite reaction may also be considered under those conditions where the CO_2 gas is able to migrate away from the place of reaction. In this case the reaction temperature depends not only on rock pressure but also on the partial CO_2 pressure at the given place. The particular case in which the partial pressure of CO_2 throughout the crust is equal to the CO_2 tension of the atmosphere is treated in another paper (in press). In this case the reaction temperature was found to decrease with increasing pressure and depth, so that wollastonite is formed, e.g., at 350° at about 30 km. depth. If the existing partial CO_2 tension in the earth's crust is less than in the atmosphere, increasing pressure and depth will depress the reaction temperature of equation (16) more steeply. However, gravitative chemical stability exists when the

partial CO_2 tension increases with depth according to the law of equilibrium (2). In this case it is more difficult to compute how increasing depth affects the formation temperature of wollastonite. However, the fact that wollastonite has a greater density than quartz or calcite indicates that, at ideal gravitational stable conditions, the temperature of formation of wollastonite decreases somewhat as depth increases.

The comparatively rare occurrence of cavities filled with CO_2 gas in regional metamorphic marbles carrying lime-silicates indicates that CO_2 commonly is able to diffuse away from the place of reaction and that the processes described above actually take place in the crust under regional metamorphic conditions.

According to the calculations of V. M. Goldschmidt (1912), who regarded wollastonite as an important geological thermometer, the lack of wollastonite in regional-metamorphic marbles should indicate a temperature lower than 500° at low pressure, and below 1,000° at high pressure. However, according to the calculations above, the lack of wollastonite in most regional-metamorphic marbles proves that the temperature has been less than 500° at low pressure and perhaps less than 400° at high pressure. This indicates that regional granitization and intense metamorphism commonly take place below 400°–500°. This is in harmony with the steadily accumulating data (Ingerson, 1937) pointing toward low temperatures of granitization and related phenomena, temperatures that are lower than the solidus point of a hydrous granitic magma (Goranson, 1931).

N. L. Bowen (1940) has treated several carbonate-silicate reactions from the viewpoint that CO_2 is unable to diffuse away from the place of reaction. Consequently, he found the higher the rock pressure, the higher the temperature required for the formation of the silicates. However, the transformation temperatures of the reactions treated by Bowen must also be more or less independent of pressure and depth, or perhaps decrease with pressure, because CO_2 is actually able to diffuse through the solid rocks.

GRAVITATIVE STABILITY OF CHEMICALLY INERT AND IMMISCIBLE CONDENSATES

Let us consider the case of two solids, *A* and *B*, as immiscible and chemically inert phases. The density of *A* is less than that of *B*. It is easy in this case to prove that the chemically stable state is established if the light *A* phase rests on the more dense *B* phase (Ramberg, 1944*a*). If *B* rests on *A* or if the two phases are mechanically mingled, then chemical forces exist which tend to diffuse *A* atoms upward and *B* atoms downward until the stable arrangement is reattained.

However, if the power of diffusion of the *A* and *B* atoms differs, then mechanical or bodily motion of the *A* or *B* crystals will also take place. If, for instance, *A* migrates upward more rapidly than *B* diffuses downward, then *B* crystals, or the whole *B* layer, sinks down through the *A* phase at the same time that *A* atoms diffuse up through the sinking *B* layer. On the other hand, if *B* diffuses more rapidly than *A*, then the downward-diffusing *B* atoms consolidate at the bottom of the system, and the expanding crystalline *B* aggregate elevates the overlying *A* crystals because of the force of crystallization of the substance *B*.

This theoretical example becomes of geological interest if we substitute for the crystalline *A* and *B* substances the light granitic and the heavier gabbroic rocks, respectively. Dense gabbroic rocks rest-

ing on light granitic rocks or magma represent a chemically unstable arrangement, so that the atoms, ions, or molecules of the granitic minerals or melt become activated and tend to diffuse up through the gabbroic rocks, which, owing to low diffusibility of the constituent elements, sink bodily.

Thus we see that the direction of the driving force of radial diffusion in the earth is in complete harmony with geological experience with regard to regional granitization. Pentti Eskola (1932), recognizing that the granitic shell was not an original feature of the earth but was formed subsequently by radial motion of substances, explained regional granitization by upward flow of light, granitic pore "magmas" through confining rocks of greater density. However, in several cases it is impossible to assume that hydrous granitic pore liquids have been present during granitization because the existing temperatures have been below the liquidus field of hydrous magmas. In other cases, also, a motion of such liquids through the very minute pore system of solid rocks is impossible. Nevertheless, field observations prove that granitic material has wandered through the solid rocks, altering them into successively more granitic types. In such cases it is important to remember that potential chemical forces are induced by mechanical instabilities and, consequently, that diffusion will account for the transportation of substances.

SUMMARY AND CONCLUSIONS

In this paper it is shown that forces exist which tend to cause elements to diffuse in directions consistent with the present radial distribution of those elements in the earth. Consequently, one of the main objections to the theory of radial diffusion in the solid crust is annulled, viz., that the driving force of diffusion is not parallel to the direction along which matter actually has been transported. P. Niggli (1941, pp. 17–18), for example, sought these "mysterious" forces when he criticized the view of Van Bemmelen (1935), who unfortunately had only a vague impression of the forces underlying diffusion.

With reference to the velocity of diffusion through solid rocks, we refer the reader to the several geological and geochemical phenomena which definitely prove that matter is able to diffuse over shorter or longer distances throughout the crust.

The influence of gravity on the vertical diffusion and the large-scale metamorphic differentiation of our globe may be summarized as follows:

1. In a homogeneous phase (mixed crystal, magma) elements with greater fictive density (p. 451) than the density of the phase will be gradually concentrated at greater depths. Elements with smaller fictive density will be concentrated at higher levels.

2. In heterogeneous systems with different phases, elements with great fictive density are commonly discontinuously enriched downward, and less dense elements are discontinuously enriched toward the top, the several phases being arranged in concentric layers in the crust.

3. Interionic bonding forces, however, will disturb this gravitative arrangement. Elements with small fictive density but strong attraction to dense phases, for example, may be absorbed in this phase at great depth. On the other hand, elements with high fictive density but with strong attraction to less dense phases may concentrate in these light phases at the top of the crust.

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THE FORM AND STRUCTURAL FEATURES OF APLITE AND PEGMATITE DIKES AND VEINS IN THE OSI AREA OF THE NORTHERN PROVINCES OF NIGERIA AND THE CRITERIA THAT INDICATE A NONDILATIONAL MODE OF EMPLACEMENT[1]

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ABSTRACT

Numerous dikes and veins of aplite and pegmatite in the Osi area of Nigeria were studied with a view to ascertaining the mechanism of emplacement. Offsetting of the invaded rocks of a magnitude and direction appropriate to dilation is so rare as to appear to be quite fortuitous. Where offsetting is not entirely absent, it is commonly of a different magnitude and frequently in the opposite direction from that to be expected on dilation of the invaded rocks. In such instances emplacement has evidently occurred along planes of shearing formed under compressional stress.

Nondilational emplacement is also clearly indicated in the case of the irregular bodies and ramifying networks of veins. The criteria advanced are similarly of significant application in the mode of emplacement of "ptygmatic folds."

Processes of metasomatic replacement, conveniently termed "aplitization" and "pegmatization," are believed to have been those responsible for the development of the bodies described, while internal evidence for the operation of such processes is sometimes provided by the textural features of the dikes themselves.

The broader petrogenetic implications emphasize the need for a closer scrutiny of the criteria that are commonly accepted as indicative of intrusive relations.

INTRODUCTION

Goodspeed (1940) showed that two categories could be recognized among dikes, depending on their mode of emplacement by dilation or replacement of the host rocks. The textural and structural features by which the differing modes of emplacement may be recognized were summarized (pp. 194–195).

Among the principal field criteria discussed by Goodspeed is the evidence provided by the oblique intersection of two dikes (pp. 189–191). Emplacement by dilation of the later of the two dikes is indicated where the earlier of the dikes shows offsetting of appropriate magnitude and direction (fig. 1, *A*). Where such offsetting is absent and the disconnected portions of the earlier dike are in complete alignment, the possibility that the later dike is of dilational type is exceedingly remote and demands the postulation of a lateral displacement that precisely corresponds to the expected offsetting. The intersection of dikes without offsetting is regarded by Goodspeed as incontrovertible evidence for a replacement origin of the later dike (fig. 1, *B*); yet such intersections have been figured without comment on the mechanical difficulties involved in a dilation interpretation. An interesting illustration in this respect is figure 10 in *The Geology of Lizard and Meneage* (Flett, 1946), which shows a dike of epidiorite obliquely cutting a dike of gabbro pegmatite, which lies within a band of gabbro schist that traverses serpentine. The epidiorite also traverses a small boss of troctolite. Not one of these intersections shows any sign of offsetting. It is clear, nevertheless, that no mechanism other than dilation and injection has been envisaged for the emplacement of the several bodies; but it is only fair to observe that the

[1] Published by permission of the Director, Geological Survey of Nigeria. Manuscript received May 3, 1948.

figure is essentially diagrammatic and is primarily intended to illustrate age relationships.

APLITE AND PEGMATITE DIKES IN THE OSI AREA OF NIGERIA

During the course of recent field studies in an unusually well-exposed area of basement complex in Nigeria, the writer examined numerous aplite and pegmatite dikes and veins. In outline, the geology of the area[2] consists of banded gneisses, granitic gneisses, and subordinate biotite- and hornblende-schists which have been involved in two granite cycles. The first was expressed as granitization *in situ* and produced a foliated granite-gneiss: during the later cycle a large body composed predominantly of coarse-grained porphyritic granite was emplaced, although in this case only part of the mass represents granitization *in situ*. A number of sheeted bodies, now consisting of metagabbro, invaded the area between the two granite cycles.

Apart from numerous semiconformable, medium-grained leucocratic veins, many of which undoubtedly antedate even the granite-gneiss, dikes and veins of both aplite and pegmatite were formed in great abundance during both the granite cycles. In summary these may be grouped as follows:

7. latest pegmatites
6. latest aplites
5. first postporphyritic granite pegmatites
4. first postporphyritic granite aplites
3. preporphyritic granite pegmatites
2. postgabbro aplites and pegmatites
1. postgranite-gneiss aplites and pegmatites

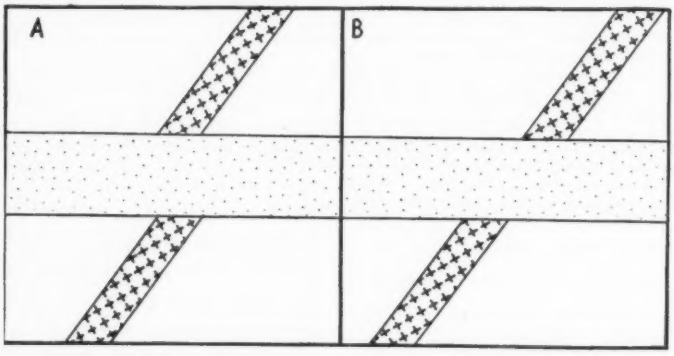

FIG. 1.—Diagram illustrating the oblique intersection of dikes. In *A*, offsetting of the earlier dike takes place, indicating a dilational mode of emplacement of the later dike. In *B*, no offsetting occurs, so that emplacement of the later dike must have taken place without dilation of the dike walls.

Figures 2, 3, 4, and 5 illustrate typical field characters of these groups of aplites and pegmatites. The absence of offsetting at intersections appropriate to emplacement by dilation is to be observed in a number of instances (figs. 3 and 4, *A*, *B*, and *C*). In addition, especial attention is directed to the following phenomena: (*a*) the commonly highly irregular form of the bodies, with "pinches" and "swells" that produce no effect on the structure of the adjacent host rock (fig. 2, *C*, etc.); (*b*) the enclosure of undisturbed patches of the invaded rock, in many places so extensive that they occupy the greater part of the width of the

[2] A detailed account of the geology is in preparation for publication by the Geological Survey of Nigeria.

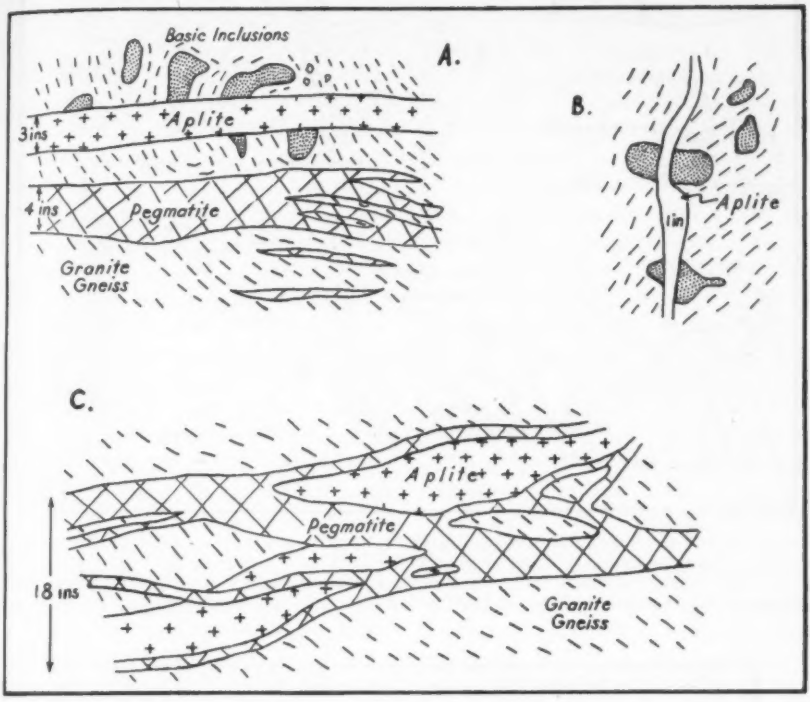

FIG. 2.—Aplites and pegmatites invading granite-gneiss, from Osi area of Nigeria. Note irregular form of pegmatite dikes and the absence of corresponding parts of basic inclusions.

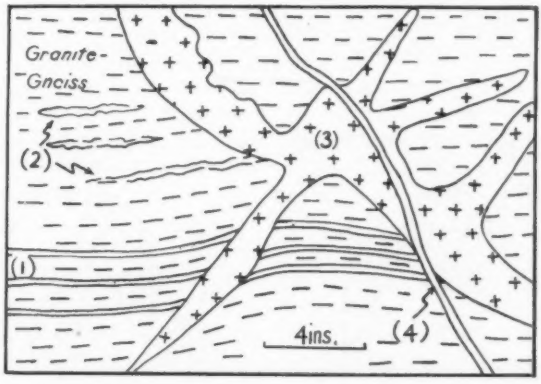

FIG. 3.—Aplites and pegmatites invading granite-gneiss, from Osi area of Nigeria. *1* represents bands in original banded gneiss, which have been partly "rejuvenated" during granitization. *2* shows streaks in granite-gneiss, due to extreme felspathization, associated with granitization. *3* represents aplite related to latest phase of granite-gneiss cycle. *4* shows pegmatite veins related to later porphyritic granite cycle.

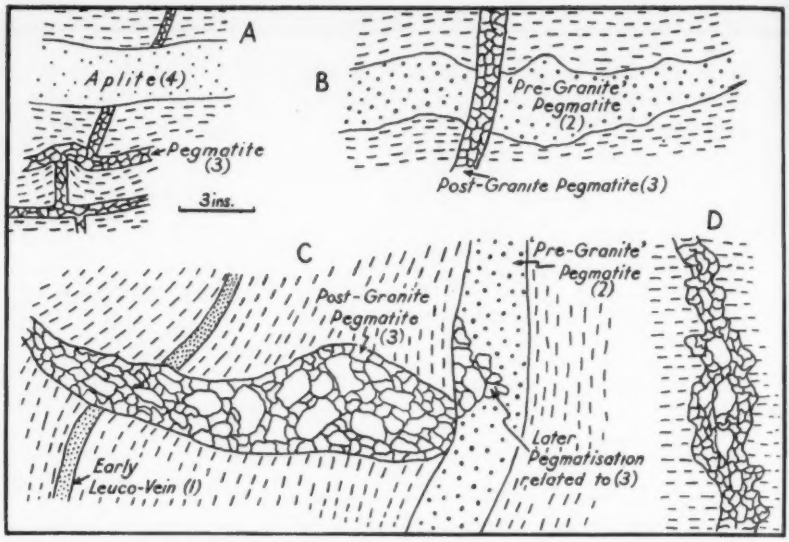

FIG. 4.—Relations between aplites and pegmatites in granite-gneiss, from Osi area of Nigeria. The pegmatite in *A* shows displacement of greater magnitude and in the wrong direction from that required by dilation of the walls of the aplite. *D* shows the character of a pegmatite transverse to the foliation in the granite-gneiss and its evident replacement habit. All dikes belong to the porphyritic granite cycle.

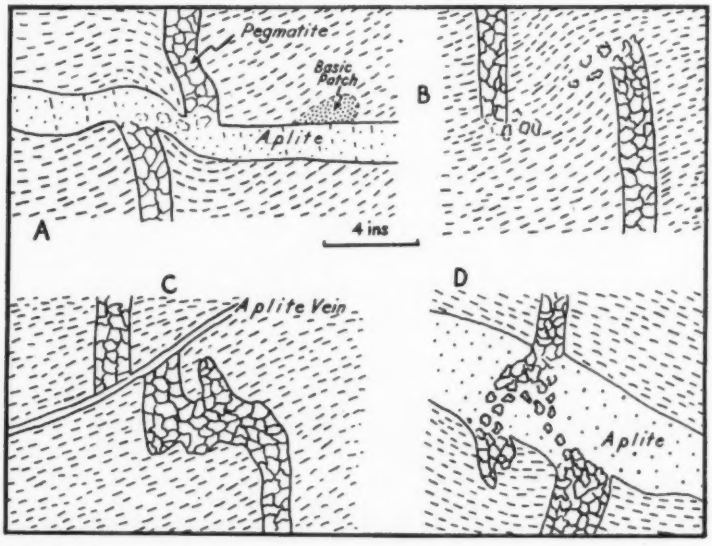

FIG. 5.—Features of aplites and pegmatites belonging to the porphyritic granite cycle and invading granite-gneiss. In all diagrams there is evidence of lateral displacement affecting the pegmatite dikes or the fissures to which they were related. In *A* and *B* there is evidence of compressional stress leading to plastic distortion of the granite-gneiss foliation. Offsetting caused by displacement is partly bridged by pegmatization of the intervening host-rock.

dike (fig. 2, *A* and *C*); (*c*) the transition of regular dikes into a series of tongues (fig. 2, *A*) or into a body with indefinite margins; and (*d*) the truncation of inclusions and basic patches in the host rock and the absence or lack of correspondence of related portions within the dike or on the opposite wall (figs. 2, *A* and *B*, and 5, *A*).

The inadequacy of dilation even as a partial mechanism for the production of the observed features cannot be overemphasized: its application in the case of the "star-shaped" body depicted in figure 3 would be particularly improbable; and, moreover, it is to be borne in mind that the addition of the third dimension serves further to intensify the difficulties that are already apparent in the application of a dilation mechanism to the present two-dimensional illustrations.

A number of the diagrams indicate the operation of a tectonic process that is incompatible with the physical environment within which dilation is likely to be associated. This is the development of shearing effects in the rocks in response to regional compressional stress. It is not proposed here to detail the evidence for the operation of such stress on a regional scale, since it is intended that this aspect will form the subject of a later contribution. In the present account attention will be confined to the relation between stress movements and the mode of emplacement of the dikes.

Observation of numerous examples of displacements of earlier by later dikes showed that the number of cases in which the displacement is of such magnitude and direction as to accord with the offsetting appropriate to dilation are so rare as to suggest that they are quite fortuitous. The actual displacements found, by comparison with those required by dilation, are generally too great, usually in the opposite direction; are unrelated to varying dike width; and commonly vary not only in magnitude but even in some cases in direction, along different parts of the same dike. Figure 4, *A*, shows that shearing movement, which apparently determined the trend of the aplite dike, has been responsible for the displacement of the earlier pegmatite. In figure 5, *C*, the relative positions of the offset portions of the pegmatite dike show that displacement occurred in response to compressional stress. The shear plane itself became the locus of an aplite vein. Precisely similar relations may be observed in displacements of structural features in the host rock, where these are nonconformably traversed by dikes.

Although it is apparent that in many instances the host rocks have undergone brittle displacements as a result of shearing stress, with the formation of fractures that have commonly been the loci along which aplite or pegmatite dikes and veins were subsequently emplaced, it is interesting to observe that in other cases the rocks have undergone plastic deformation. Indications of the latter are seen where the relative lateral displacements on either side of a dike show rapid variations both in magnitude and in direction. No simple mechanical explanation of such features is possible, for the rocks have behaved as if they were capable of localized compression and elongation.

Related features are seen in figures 4, *A*, and 5, *A*, where, near the intersection of two dikes, the invaded granite-gneiss has been affected by what may be described as "plastic rotation" along an axis perpendicular to the plane of the diagrams. In figure 5, *A*, especially it is apparent that the gneiss underwent rotation that is relatively greater than that

shown by the aplite dike, a feature which suggests that deformation of the gneiss commenced before emplacement of the dike. The pegmatite (or its predetermining fracture) has experienced a corresponding displacement on either side of its intersection with the aplite. Similar plastic deformation induced by compressional stress operating at right angles to the plane of foliation is also demonstrable in figure 5, *B*. Dislocation has been distributed through a zone and is not confined to a definite shear plane. As in *A* and *C* of the same figure, compression is indicated by the "overlapping" of the displaced ends of the dike.

Figures 6 and 7[3] are of interest in that they depict "anomalous" features of otherwise regular and persistent large aplite dikes. Figure 6 shows part of a large dike, which, were it not for the fact that it is traversed in one place by an offshoot of the host rock, would be regarded as having been emplaced within the latter. Actually, the formation of the dike preceded that of the surrounding granite. The field evidence demonstrates that the dike was emplaced within the coarse porphyritic granite which forms the greater part of the granite mass. Later this granite was locally replaced by a more even-grained leucocratic granite. Relics of the earlier granite are now represented only by inclusions within the aplite dike, which at one point (not shown in diagram) consists merely of narrow septae and marginal selvages enclosing undisturbed blocks of the earlier granite. The unrotated character of the inclusions is apparent from the conformance of the characteristic "flow-layering" of the feldspar phenocrysts.

The features illustrated in figure 7 would place quite extraordinary demands on a hypothesis of dilation with concomitant injection of aplite. Both apophyses are continuous with the same larger dike, yet one obliquely cross-cuts the other without effecting displacement.

[3] The original photographs from which figs. 6 and 7 were prepared will be reproduced in the publication by the Geological Survey of Nigeria.

ANALOGOUS FEATURES FROM OTHER AREAS

In an account of the Cnoc nan Cuilean area of the Ben Loyal complex of Sutherland (King, 1943), the writer employed two diagrams (figs. 1 and 2) illustrating dikes and veins of aplite and pegmatite which traverse the variable syenites of the marginal zone of the body. These diagrams have been redrawn as figures 8 and 9 in this paper. At the time of their original publication, the writer observed nothing in the mutual relations of the dikes which might reflect on their mode of emplacement. It is now apparent that they are beautiful examples of the absence of offsetting of earlier by later obliquely intersecting dikes. Furthermore, the absence of corresponding portions of truncated "basic patches" from the opposite sides of small dikes are even more clearly shown than are similar features from the Osi area. The significance of parallel sides to undulating veins will receive attention later (p. 469).

Comparable features in aplite dikes were also recorded by the writer during the course of mapping in the Kawungera area of central Uganda during 1945, although it is only in the light of the examples from the Osi area that the significance of the Uganda occurrences has been appreciated. In the Kawungera area a mass of coarse-grained porphyritic granite invades phyllites, sandy phyllites, and quartzitic sandstones of the Karagwe-Ankolean system but is older than the Singo series, which here consist of a thin basal sandstone followed by

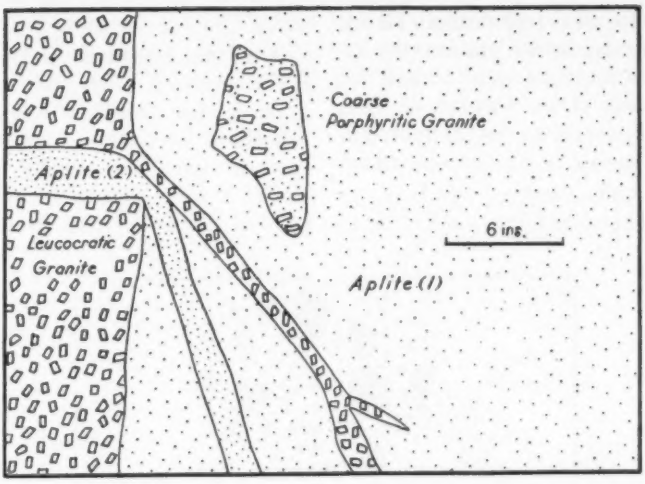

FIG. 6.—Part of a large aplite dike within coarse leucocratic granite: a late phase of the porphyritic granite. For interpretation see text, p. 464. From Osi area of Nigeria. (Drawn from a photograph.)

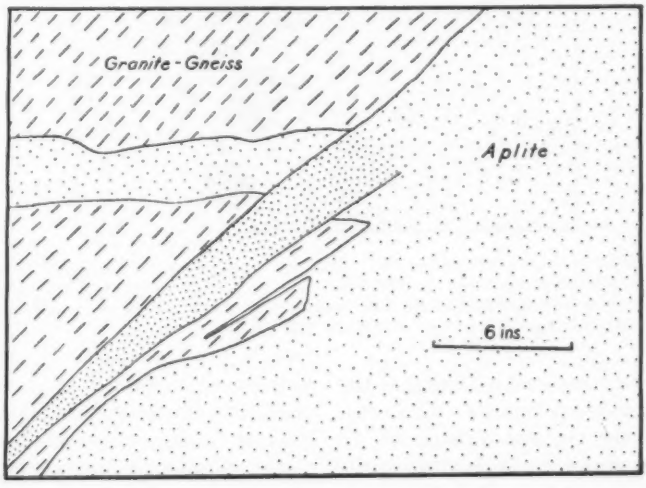

FIG. 7.—Margin of large aplite dike cutting granite-gneiss, showing mutual relation between two apophyses. Although both pass continuously into the main dike, one is cross-cutting toward the other. From the Osi area of Nigeria. (Drawn from a photograph.)

silicified "siltstones" (King, 1947, pp. 38–40). Aplite dikes and veins and a few pegmatites traverse the granite; the accompanying diagrams (figs. 10 and 11) illustrate some of the features of these aplites. In figure 10, *A*, the partly irregular form of the aplite dike and associated veins, the main trend of which is apparently determined by parallel fractures, may be compared with that of the composite aplite-pegmatite dike (fig. 2, *C*) from the Osi area. Figure 10, *C*, provides an example of offsetting in the opposite direction from that demanded by dilation of the dike walls. In figure 11 it is clear that, if the smaller body had been emplaced by dilation-injection, the width of the larger dike would have been the same on both sides of the later body.

RAMIFYING NETWORKS OF DIKES OR VEINS

A notable feature of Archean terrains is the presence of irregular networks of anastomizing dikes and veins of aplite, pegmatite, or quartz. The sheets may conform partly or wholly to certain trends or may be entirely irregular, depending, apparently, on the presence or absence of directional structures within the host rock along which planes of weakness or fracture may develop preferentially. Similar bodies are also found, though generally more sporadically, in relation to granite plutons of later ages.

Little attention appears to have been paid to the mechanism by which the host rocks have accommodated themselves to the emplacement of such complex bodies,

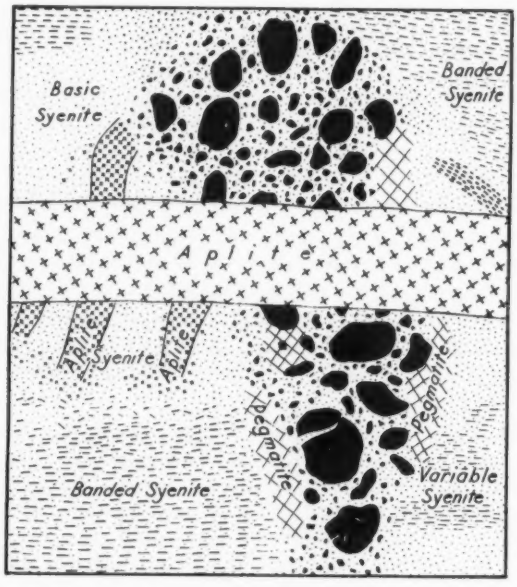

FIG. 8.—Aplites in "variable" syenites from Cnoc nan Cuilean area of Sutherland. The large aplite (which is about 4 inches across) truncates "basic patches" and earlier aplites; but corresponding portions are not represented on opposite sides of the dike. Areas of diffuse aplitization and pegmatization may also be observed. Redrawn from Fig. 1 of Vol. 98, Quart. Jour. Geol. Soc. for 1943.

which commonly belong not merely to one but to a number of epochs. Figures 12 and 13 illustrate typical assemblages of such veins from the Osi area. Figure 9 presents similar features from another region. Attention is directed to the following characters: (*a*) the persistence of many of the veins and their comparatively constant width, despite numerous intersections and the numbers of different trends that are represented; (*b*) the many examples of mutual intersections without offsetting of either vein; (*c*) the continuity of structure in the various isolated blocks of host rock; and (*d*) the parallelism of the walls of those veins that have undulating forms as well as of those that are straight.

In order to demonstrate the effect on host rocks of uniform tensional stress necessary to accommodate a typical network of veins, figures 14 and 15 were drawn to illustrate the relative positions of blocks of host rock both before and after uniform dilation has operated along a given set of fractures. The following features of the diagrams are to be noted: (*a*) the longer and straighter fractures do not dilate in such a fashion that they could be occupied by regular veins of constant width; (*b*) intersections almost invariably show offsetting of a magnitude that depends on the relative size and shape of the adjacent blocks; (*c*) formerly continuous structures in the host rock become displaced; and (*d*) precise parallelism of the walls of undulating fractures is never maintained after dilation—the disparity increases with increasing curvature of the vein.

It is readily apparent that the resultant picture in no respect accords with the features actually observed in the field. Comparable patterns may perhaps be confined to examples of brecciation that has occurred at relatively shallow depth. In particular it should be noted that a dilation mechanism is almost incapable of producing tongue-like apophyses which do not completely traverse isolated blocks, such as are almost invariably found in actual examples of vein networks (fig. 12).

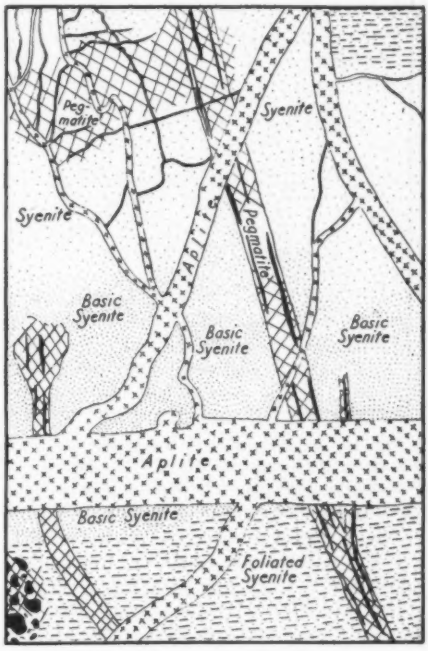

FIG. 9.—Network of aplite and pegmatite dikes and veins in "variable" syenites from the Cnoc nan Cuilean area of Sutherland. Numerous intersections occur without offsetting, while the general pattern of the ramifying veins is not that found with dilation (see also p. 465 of text). Redrawn from Fig. 2 of Vol. 98, Quart. Jour. Geol. Soc. for 1943.

The general principle of the above demonstration is not affected if we consider the tensional stress to predominate in one direction; in such cases a square of the original pattern will be rectangular or rhomb-shaped after dilation, whereas, even if the host rocks are regarded as capable of nonuniform dilation, compen-

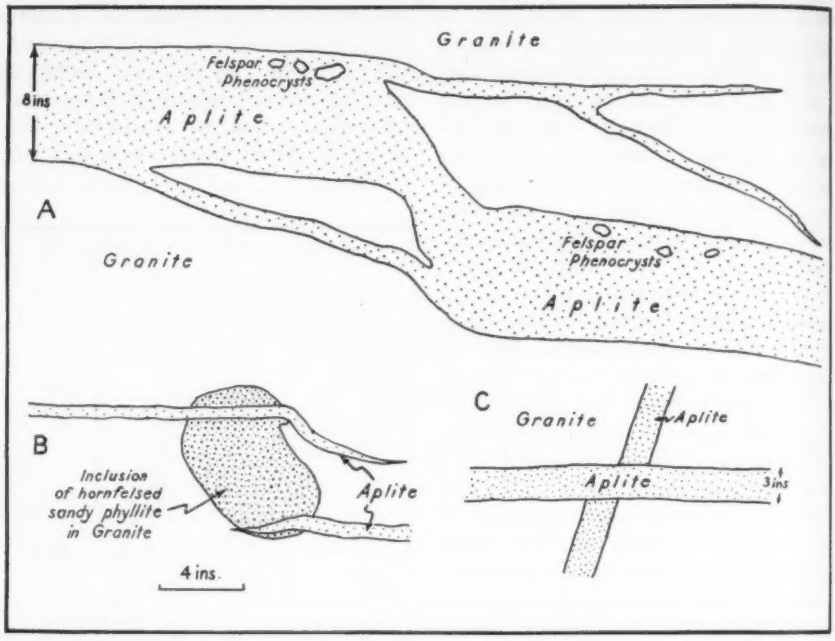

FIG. 10.—Aplites from the Kawungera granite area of central Uganda. *A* shows the development of a dike along two parallel fractures. *B* illustrates a similar feature on a smaller scale, in which aplite veins cut an inclusion in granite without displacement of the walls. *C* shows offsetting of an earlier dike by a later of larger magnitude and in the opposite direction from that expected to result from dilation.

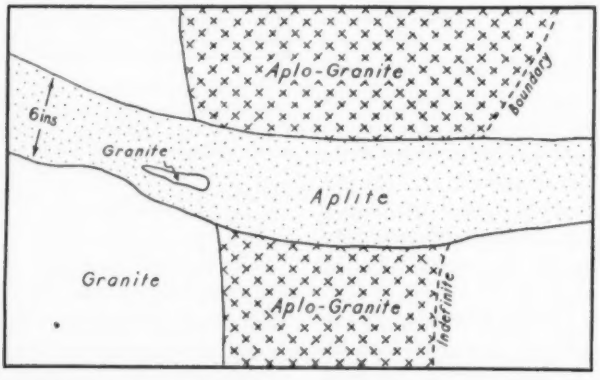

FIG. 11.—Aplite and aplo-granite, cutting granite in the Kawungera area of central Uganda. The aplite dike has evidently developed without dilation of the host-rock.

sation of the numerous local irregularities must occur over larger units. Again, it may be noted that, whether the dikes and veins belong to one or to more than one period of formation, the general argument remains the same.

partings only at the crests of the plications, whereas those portions of the fracture that are to be occupied by the subparallel limbs of the vein folds will not open at all.

Read (1928) and, more recently, Bud-

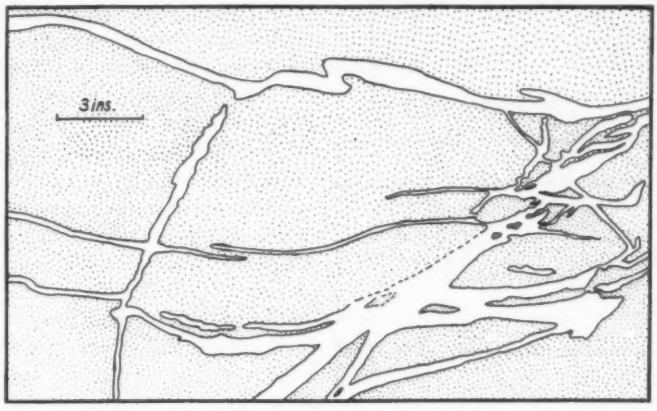

FIG. 12

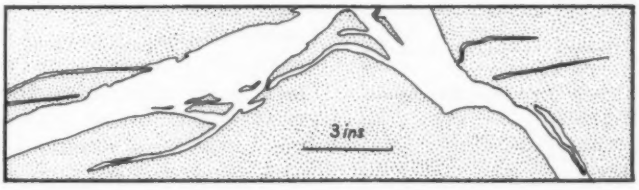

FIG. 13

FIGS. 12 and 13.—Examples of ramifying networks of veins of aplite composition and texture which invade partly granitized metagabbro in the Osi area of Nigeria. The margins of the veins are often indistinct, while diffuse patches of similar composition also occur in the host-rock. (Drawn from photographs.)

PTYGMATIC VEINS

It is readily apparent by experiment that a vein will not have parallel sides if it has been emplaced by simple injection between the dilated walls of an undulating fracture. In the case of ptygmatic veins the difficulty is especially acute, since it is easily verifiable that dilation in a general direction at right angles to the main trend of such a vein will cause

dington (1939, pp. 164–165) state that ptygmatic veins have their original form and were never planes. Read (p. 75) in a number of sketches amply demonstrates this contention—a conclusion that is the more incontestable because it is evident that the foliation in the invaded Moine series largely agrees with the original sedimentary bedding. Nevertheless, Read maintains "the clearly intrusive

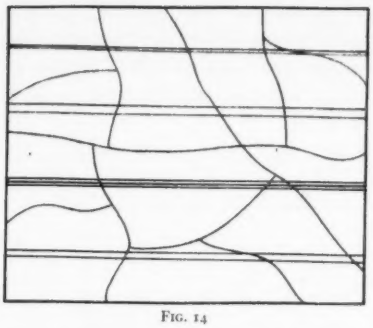

FIG. 14

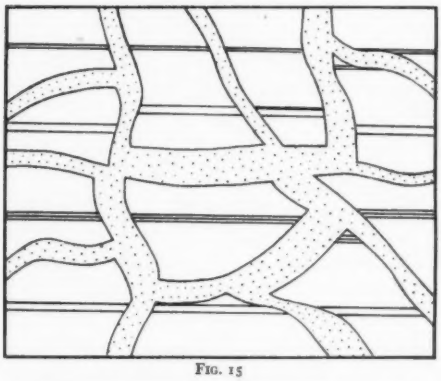

FIG. 15

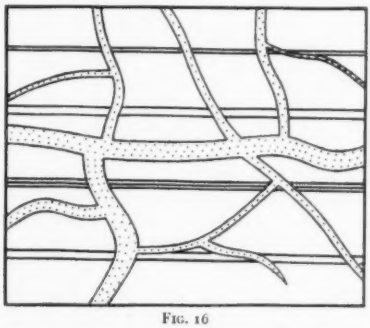

FIG. 16

FIGS. 14–16.—Fig. 15 illustrates the effect which may be expected from uniform dilation along the fractures indicated in Fig. 14. Noteworthy features are the displacements of original structures and of intersections; the lack of uniform width to channels following the longer fractures; and the absence of parallel sides to channels along curving fractures. In Fig. 16 replacement is imagined to have operated along the same set of fractures. The correspondence between the features so produced and those found in field occurrences of networks of dikes and veins is to be noted.

igneous nature of the veins, and their definite association with great igneous bodies of unquestionable origin." The tortuous character of the veins is ascribed to the "absence of easily opened plane channels for the injected material."

The evidence presented in this paper, in its application to ptygmatic veins, leads clearly to the conclusion that, if their mode of emplacement was by dilation-injection, their present form must have been attained by deformation of original flat sheets; whereas, if it can be proved that the latter was not the case but that their present form is an original character, then the veins have not been formed by dilation with concomitant injection.

The writer, although not contesting the evidence that ptygmatic veins, in the examples given by Read, owe their form to an original character, has noted examples in which similar bodies have evidently been produced by deformation of sheets of simpler form. In figure 17 the close conformity of many of the tortuous veins with relic structures indicates that they were originally semiconformable veins in the banded gneisses, but the axes of the minor plications now accord with the impress of a later deformation. "Rejuvenation" of the earlier veins and further accretion of leucocratic material have taken place during the granite-gneiss cycle, to which the larger irregular patches are to be solely ascribed. The two more regular aplite dikes seen in the diagram are related to a closing phase of the later cycle, and their general conformability to the granite-gneiss foliation is apparent.

THE MODE OF EMPLACEMENT OF NON-DILATIONAL DIKES AND VEINS

Because it is demonstrable that the form displayed by the dikes and veins which have been described above could

not have resulted from the dilation of fracture walls accompanying the injection of the dike-filling material, it is apparent that the volumes now occupied by the dikes and veins were formerly occupied by host rock. The latter must therefore have undergone replacement or displacement. Whereas, in the case of larger bodies, it is not necessarily easy to preclude the possibility that the invaded rocks have disappeared by a process of disintegration and mechanical removal, such as "stoping" (Billings, 1925), for small and irregular dikes or ramifying networks of veins this is virtually impossible. The vein networks are often essentially "mechanically closed" systems and may be completely so in the two dimensions in which they can normally be studied. In some cases, too, the veins and host rock are of strongly contrasting composition (such as quartz-feldspar aplite within biotite- or hornblende-schist). The usual absence of signs of "contamination" is inconceivable if bodily removal of the invaded rock is postulated. The transformations must, therefore, have been effected in terms of ultimate crystal units by processes of metasomatic replacement, which, in order more specifically to imply the uniformity of the resultant products, may be designated as "aplitization" or "pegmatization."

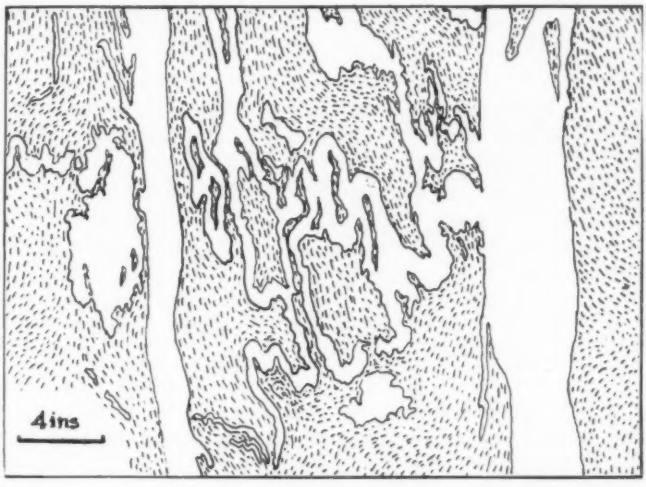

FIG. 17.—Complex system of veins in partly granitized banded gneisses. The veins are essentially of aplite but are somewhat pegmatitic in places and are largely based on semiconformable leucocratic bands in the earlier gneiss, which were crumpled and "rejuvenated" during the granite-gneiss period of deformation. The original conformable structure is apparent where the relic foliation of the banded gneisses is largely retained (see especially lower center). Later granitization is tending to produce a new foliation parallel with the axes of the minor folds: this direction is also followed by the two dikelike bodies of aplite which belong to the later phases of the granite-gneiss cycle. The development of the larger aplite areas is to be ascribed to the later cycle, while "rejuvenation" has caused the dikes to appear as continuous with the contorted veins. From the Osi area of Nigeria. (Diagram drawn from a photograph.)

It is clear that the aplitizing or pegmatizing agencies operated initially along fractures or other planes of weakness within the host rocks. In strongly banded or foliated rocks semiconformable veins are commonly found, but the larger dikes tend to be disposed along shear planes in areas where these are present. More irregular dikes and veins form if the host

rock is deficient in preferred directional structures or if, in banded or foliated rocks, the dikes and veins have formed in directions that do not accord with natural planes of easiest fracture. Highly irregular margins, evidently ascribable to the effects of uneven replacement, are commonly characteristic of such bodies, of which examples may be seen in figure 4, *D* (pegmatite formed across the foliation of granite-gneiss), and certain of the veins in figure 12 (weak development of preferential directions in the host rock).

processes must have produced such contacts have been described (Webb, 1946). As a further example may be cited a small dike of pegmatite which cuts porphyritic granite in the Osi area. The dike contains phenocrysts of microcline which are indistinguishable from, and possess the same orientation as, those in the adjacent granite. At the contact of the dike, certain of the phenocrysts are shared by both granite and pegmatite, despite the fact that in other respects the junction is perfectly sharp (fig. 18). The conclusion

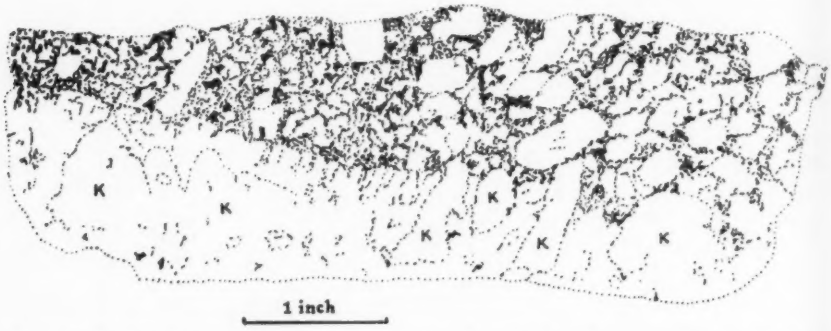

FIG. 18.—Junction between quartz-felspar pegmatite and marginal porphyritic granite from the Osi area of Nigeria (*K* 8). The microcline phenocrysts of the granite are also found within the pegmatite: some are actually shared by both rock types. In the pegmatite, further accretion of microcline has occurred, with the production of larger, more rounded plates. The principal areas of microcline in the pegmatite are indicated (*K*), but usually enclose small blebs of quartz and oligoclase.

In figure 16 a network of veins has been drawn on the assumption that they were formed by agencies of replacement which operated uniformly from the same set of fractures that were employed in figures 14 and 15. It is readily apparent that the resultant pattern and individual features of the veins accord with those that were summarized on page 466 as characteristic of networks of veins of the type under discussion.

The presence of "sharp" contacts has often acted as a deterrent toward explanations that depend on replacement, although examples in which replacement

that the pegmatite has formed essentially by a modification of the matrix of the granite in this case appears inescapable.

Those who concede that replacement may be partly responsible for the formation of dikes and veins such as those described and who retain dilation-injection as the primary mechanism in dike formation must account for the formation of a uniform rock by two entirely different processes. Moreover, they must recognize that replacement, which predominates in the marginal portions of the bodies, is responsible for the sharp contacts.

Under the microscope, even the sharp contacts that in the field are characteristic of the more regular aplite and pegmatite dikes and veins are no longer so distinctive. Instead are found the "crystal continuities" that Stillwell (1918, p. 194) regarded as indicative of "metamorphic diffusion." Individual crystals are shared at all points along the contact by both dike and host rock. In spite of Stillwell's evident confidence that "igneous" phenomena are ascribable to an entirely different mechanism, it is significant that he believed that such metamorphic diffusion took place in the solid state (1918, pp. 12, 204–208). Reynolds (1947, p. 221) has recently suggested the possibility that "sharp" contacts represent diffusion limits, and she draws attention to analogous features in experimental results on solid diffusion.

In all the areas from which examples have been cited, besides discrete dikes and veins with "sharp" contacts, there also exist less regular and more indefinite patches of aplitization and pegmatization (figs. 9 and 17). Even along a single body, variations from sharp to indistinct boundaries are not uncommon. The retention of relic structures of the host rock, apparent in the hand specimen, is a common feature of the irregular patches of aplite and pegmatite, and, under the microscope, even the regular bodies often betray an origin by replacement of the pre-existing rock. This is especially evident in certain pegmatites, wherein the matrix shows close textural and mineralogical affinities with the invaded gneisses or granite. A detailed statement of the petrographic evidence for this contention will be presented in the full account of the Osi area.

In the development of the dikes and veins under discussion, introduction of material on a considerable scale has not normally been involved. In general, the aplites and pegmatites are richer in silica, alkalies, and alumina and poorer in the basic constituents than are the gneisses or granites that form the host rock. The destination of the displaced basic constituents is not always a matter of conjecture because, most especially in the case of the irregular masses and the semiconformable veins, basification of the host rock in the vicinity of the contacts is commonly observed (fig. 19). Precisely similar features were recorded by Read, who describes "biotite selvages" to the veins of the "injection-gneisses" of Sutherland (1931, fig. 9 and compare also pl. II, *A*). A comprehensive study of the geochemistry of analogous, but larger-scale, phenomena in relation to granitization has recently been made by Reynolds (1946).

FURTHER PETROGENETIC CONSIDERATIONS AND CONCLUSIONS

It has been demonstrated in the foregoing account that sharp contacts and transgressive relations with the host rock are by no means adequate criteria for the assumption that dikes and veins have been emplaced by a dilation-injection mechanism. In marked contrast to minor bodies that have been emplaced at higher crustal levels, aplites and pegmatites show nothing to correspond with the obvious chilled margins that characterize, for example, basic dikes. In the absence of such incontrovertible evidence of a former magmatic nature, it is necessary, therefore, to examine closely the relations between dike form and structure of the country rock before rejecting a replacement hypothesis in favor of one of dilation with concomitant injection.

The criteria on which such discrimination is possible have significant applications besides those which relate to the

mode of emplacement of dikes and veins. As an example may be cited the illustration (fig. 99, p. 271) in *Igneous Rocks and the Depths of the Earth* (Daly, 1933), which is entitled "Arrested Stoping at the Roof of the Lausitz Granite Batholith." The mechanism implied by this process is clearly indicated by Daly in the text: the blocks of country rock are inevitably be found where dilation has occurred. In particular, it may be noted that the corresponding corners of adjacent blocks, which also represent the terminations of apophyses, are in precisely the positions that they would occupy were the apophyses not there at all. The small horizontal vein near the right of the figure follows a direct course which

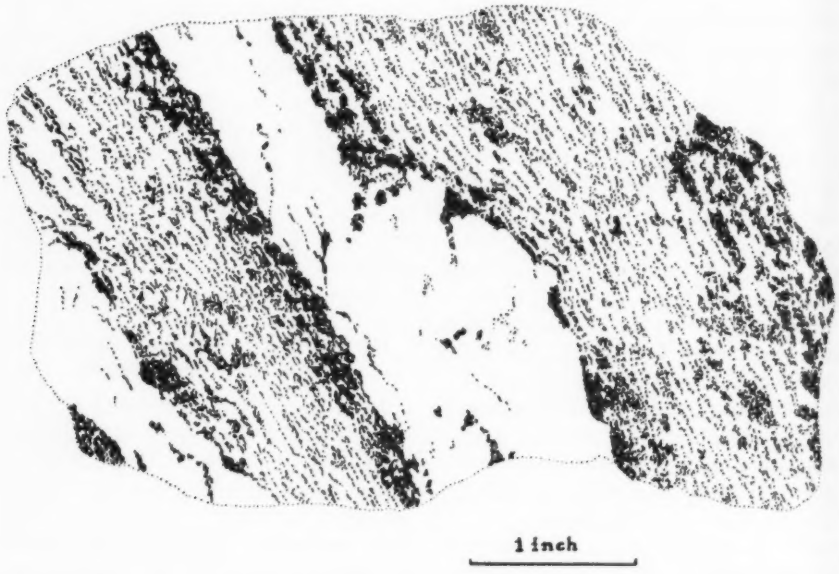

FIG. 19.—Leucocratic veins in partly granitized banded gneiss (granite-gneiss cycle) from the Osi area of Nigeria (*K* 134). Marginal basification (hornblende and biotite) adjacent to the leucocratic veins is well developed, while there is a concomitant partial disappearance of the well-marked foliation. The darker gneiss at the extreme right of the illustration indicates proximity to another vein (not shown on this side of the specimen).

regarded as having been riven apart by numerous granite apophyses and ultimately, when completely surrounded by granite magma, as becoming capable of independent relative movement.

Although it is appreciated that the details of such a diagram should not be subjected to too precise a scrutiny, it is evident that it shows none of those features which, as summarized on page 467, will neither affects, nor is affected by, its intersection with two other veins.

The relevance of this example is the more apparent if one recalls that it was the discovery of veins of granite continuous with a main granite mass penetrating the mica schists and limestones of Glen Tilt that Hutton found especially convincing evidence for the existence of granite magma (Read, 1943, pp. 67–68).

Now that the presence of sharp contacts can no longer be regarded as conclusive evidence in this direction, the examination of the form of such apophyses and their relation with the structures of the invaded rocks has become of critical importance, and, although the absence of indications of dilation does not preclude the possibility of magmatic injection, its likelihood in such instances becomes exceedingly remote.

ACKNOWLEDGMENTS.—I wish to express my thanks to Professor Arthur Holmes, Dr. Doris Reynolds, and Professor Neville George for reading the manuscript and for their many helpful comments.

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DEFORMATION OF QUARTZ CONGLOMERATES IN CENTRAL NORWAY[1]

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ABSTRACT

All three axes of deformed ellipsoidal pebbles have been measured in quartz conglomerates in the Sparagmite rocks (arkosic sandstones) of central Norway. More than one hundred measurements have been treated statistically. Shearing motion is considered as the main element of the deformation, and various problems concerning the deformation are discussed.

INTRODUCTION

In orogenic regions the movement of peripheral nappes over long distances has been well known to geologists for a long time. For geological reasons rocks are supposed to have been thrust over large distances, perhaps many hundred kilometers. Such thrust movements are accompanied by deformation of the rock itself. From a geological and microscopic examination it is easy to get a rough qualitative estimate of the degree of deformation. But it is a very difficult task to get a quantitative determination of rock deformation. The only paper that I have found chiefly devoted to this problem is that recently published by Ernst Cloos (1947). By measuring the deformation of the oöids in oölitic limestones, Cloos obtained quantitative expressions for the rock deformation in the South Mountain fold of southern Pennsylvania and Maryland.

In the present study I have tried to give a quantitative expression of the deformation of sandstones by measuring the three axes of deformed pebbles in quartz conglomerates. The results are not analogous to those of Cloos because the deformation is supposed to be of a different type than that investigated by Cloos.

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GEOLOGY OF THE AREA

The quartz conglomerates investigated are found in the Sparagmite formation in central Norway (fig. 1). "Sparagmite" is a Scandinavian designation of certain coarse feldspathic sandstones, with a feldspar content of 20–30 per cent, rarely 10–40 per cent. In the Moelv division of the formation there are some rare quartz conglomerates. The pebbles consist of pegmatitic quartz and fine-grained quartzites. The undeformed pebbles had diameters between 3 and 10 cm. The thickness of the conglomeratic layers amounts to a couple of meters.

The Sparagmites are of Eo-Cambrian age. They were deposited on pre-Cambrian gneisses and were overlain by Cambro-Silurian strata, the thickness of which probably did not exceed 2,000 meters. During the Caledonian orogeny the Sparagmites were thrust toward the southeast. The thrust width is estimated by various authors to be between 40 or 50 and 300 km.

During the deformation the Sparagmites were slightly metamorphosed, as shown by the formation of sericite, alteration of feldspar, and the formation of a marked plane of schistosity. The schistosity is parallel to the conglomeratic layers and to minor traces of primary

stratification, such as the border between a layer rich in feldspar and a layer free from feldspar or thin dark bands of layers rich in clay. Few exceptions exist. It is believed, therefore, that no general folding occurred in these rocks.

The displacement of the rocks must have taken place in the following way: northern part of the Sparagmite area, the originally spherical pebbles were deformed to triaxial ellipsoids. Here the planes of schistosity display gentle dips ($< 30°$), but the strike is likely to vary from one mountain to another. The long and intermediate pebble axes lie within the schistosity planes, and the short

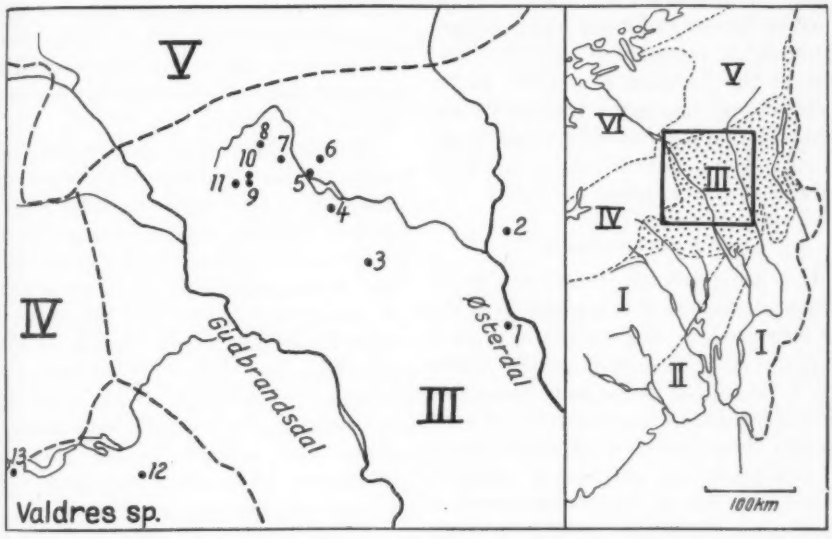

FIG. 1.—*To the right:* Key map. The formations are: *I*, pre-Cambrian gneisses. *II*, sediments (Cambro-Silurian) and igneous rocks (Permian) of the Oslo region. *III–VI*, rocks of the Caledonian orogeny. *III* (*dotted*), sparagmites. *IV*, thrust igneous rocks of Jotunheimen, etc. *V*, folded Cambro-Silurian of the Trondheim region. *VI*, highly metamorphic gneisses of the root zone.

To the left: Map showing localities of conglomerates, carrying pebbles which have been measured. (This area corresponds to the one inside the black square on the small-scale map.) *1*, Gråvola, 15 km. N.NW. of Koppang, Østerdal; *2*, Kivfjell, 19 km. N.NE. of Atna, Østerdal; *3*, Storfjellet, 19 km, W.SW. of Atna, Østerdal; *4*, Gråhøgda, 10 km. S. of Atnesjø; *5*, at Hyrsa, 2 km. NW. of Atnesjø; *6*, Setervola, 3 km. NE. of Atnesjø; *7*, Veslekollhø, Rondane area; *8*, top of Högronden, Rondane area; *9*, col between Rondholet and Storbotn, Rondane area; *10*, near top of Rondeslottet, Rondane area; *11*, Rondhalsen, Rondane area; *12*, northern border of N. Etnedal quadrangle; *13*, Eastern end of Lake Bygdin. "Valdres sp." is the so-called "Valdres sparagmite," a Caledonian flysch similar to the sparagmite.

The formation moved intermittently as a nappe on the underlying mass. This was accompanied by minor discontinuous movements along bedding planes, with the shale zones serving as a lubricant. Simultaneously, the rock itself underwent a continuous shearing movement, which was so pronounced that, in the axes are normal to the plane. Everywhere the long axes are oriented in a northwest-southeast direction, the direction of the main transport, which is marked by a pronounced lineation. This orientation of the pebbles agrees with what was found by Strand (1945, p. 17), and by Kvale (1945, p. 193).

DEFORMATION OF THE CONGLOMERATES

DEFINITIONS

In accordance with Cloos (1947, p. 853) the following designations have been used: The spherical pebbles have a radius r; after deformation the ellipsoids have semiaxes, a, b, and c. The tectonic axes *a*, *b*, and *c* are oriented as defined by Sander and many other authors. It should be emphasized that the direction of the lineation is parallel to the direction of the main transport. The same relation was found by Kvale (1945, p. 204), who considered the *a* axis to be parallel to the lineation.

THE MEASUREMENTS

It is very difficult to find pebbles in which all the three ellipsoidal axes can be measured, as emphasized by Strand (1945, p. 18). In most cases the pebble has to be cut out of the rock so as to secure a reliable measurement of all axes. In only three localities, where the planes of schistosity are extensively exposed, were more than 10 pebbles measured. In most cases it has been possible to measure with an exactness up to 1 mm., but not always, as shown by table 1.

During six years of field work I have measured 118 pebbles (table 1). In addition, ten measurements from similar rocks are taken from Strand (1938, p. 44; 1945, p. 19). The two localities of Strand are found within the "Valdres Sparagmite," a flysch horizon situated between two nappes of igneous rocks overlying the Sparagmite formation.

THE MODE OF DEFORMATION

Before treating the measured properties, it is necessary to consider the type of deformation.

The deformed rocks, described by Cloos (1947), consist of shales and limestones. These rocks were shear-folded by the orogenic movement, so that the original thickness of the beds increased. The Sparagmites, on the contrary, have been thrust without folding. The shear movement which deformed the conglomerates (and the rocks) could have acted along one or more shear planes. In a quartzite from western Norway, Kvale (1945, p. 199) found three planes of shear, one macroscopic and two microscopic. The latter planes were indicated by the orientation of the micas.

The deformed Sparagmites have one macroscopic *s* plane—the plane of schistosity. This plane is defined by a marked cleavage and by the parallel arrangement of elongated quartz and feldspar grains and of the faintly wavy zones of muscovite. A study of thin sections from the Rondane area seems to indicate that the muscovite has only one preferred orientation, which is parallel to the elongated quartz and feldspar grains and parallel to the plane of schistosity. No other macroscopic or microscopic structure is observed.

These facts suggest that the conglomerates were deformed by shear parallel to the plane of schistosity. The general case is then a triaxial deformation by inhomogeneous shear. But, in order to calculate the deformation, some simplifying assumptions are necessary.

First, the shear is assumed to be homogeneous within each layer of conglomerate. The shear movement along *a* is no doubt the principal movement, and shear in the *b* direction plays a subordinate role. Therefore, the shear may be assumed to be, as a first approximation, a pure shear in the *a* direction.

Many authors have discussed this type of deformation, and it has many names: "shearing motion (or scission)" by Becker (1893, p. 24), "Scherbewe-

gung" by Sander (1930, p. 15), "simple shear" by Nadai (1931, p. 294), and "einscharige Gleitung" by Schmidt (1932, p. 45). Following Becker, the author prefers "shearing motion."

Becker (1893) was the first to give mathematical expressions to the different types of rock deformation which might occur. But he did not deduce the equations of the shearing motion. This is done by Nadai (1931, pp. 295–297).

The shearing motion may be defined in the following way:

The rock is cut up into infinitely thin

TABLE 1

DIAMETERS OF THE MEASURED PEBBLES*

Locality	Pebble Dimensions (Cm.)			d
1	3.5	2.8	1.8	2.60
	8.3	7	3	5.59
	8.8	6.7	2.8	5.49
2	9	6	4.5	6.25
	7	7	4	5.81
	6.5	6	3	4.90
	6	3	2.3	3.46
	7	6	4.4	5.70
	8.3	4.8	3	4.93
	4.5	3.3	2.1	3.15
	6.5	4.5	2.3	4.07
	7.5	4.7	2.8	4.63
	6.5	4.8	3.8	4.91
	14	9.5	2.4	6.84
	8.5	6.2	3.5	5.69
	7.5	3	2	3.56
	9.5	7	3.5	6.15
3	18	9.7	3.5	8.49
	23	10	3	8.84
	12	5	2	4.94
	14	10.5	3.5	8.01
	12	5	1.5	4.49
	9	6	2.2	4.90
	7.5	5	1.8	4.07
	17	10	2	7.00
	6.3	4.3	1.1	3.10
	13.5	6	2	5.48
	8	4.5	2.2	4.30
	4.8	2.8	1.1	2.46
	5.3	4	1.5	3.17
	10	7.5	1.7	5.03
	11.2	5.2	1.4	4.34
	21	11	2.5	8.32
	8.5	5.3	1.4	3.98
	11	6	2	5.10
	17	8.5	3	7.56
	6.7	4.7	1.8	3.84
	6	4.5	2.3	3.96
	13.5	9	2	6.24
	10	6	2.5	5.32
	24.5	11.2	3.6	9.95
	9	6	2.5	5.13
	12.5	7	2.6	6.10
	8	6	2.2	4.73
	9.5	4.8	1.8	4.35
4	22	9	4.5	9.60
	14	9	4	7.96
	19	10.5	3.4	8.78
5	6	2.5	1	2.47
	6	3	1.2	2.78
6	12	5	2	4.94
	8	5	1	3.42
	8	6	1.5	4.16
	9	6	2.5	5.16
	8	4	2	4.00
7	20	6	2	6.22
	8.3	6	2.4	4.92
	8	4.4	2.0	4.13
	17	8	2.5	6.98
	13	8.8	3	7.00
	9.5	3.5	1.2	3.42
	7.3	4.5	1.7	3.82
	8	5	1.7	4.08
	17.5	14.5	4	10.05
	5.5	3.7	2	3.44
	12.5	9	2.1	6.18
	11	7	1	4.26
	5.5	4	1.1	2.90
	9	5.5	1.8	4.47
	8.5	6	1.3	4.05
	16	10	1.8	6.65
	6.2	4.5	1.3	3.31
	10	5	2	4.65
	15.5	12	2.5	7.75
	17	8.5	2.3	6.92
	6.5	4.5	2.5	4.18
	8.5	6	1.7	4.43
	14	8	3	6.96
	8.4	4	1.8	3.92
	8.5	6.5	2	4.82
	7	6	2.3	4.59
	7	5	2.5	4.44
	7	6	1.5	3.98
	5	5	1.6	3.42
	10	7	2.5	5.60
	8.6	6	2	4.69
	10	8	2	5.44
8	6	2.5	1.2	2.62
	6	3	1	2.62
9	30	6	0.5	4.49
	20	5	0.3	3.11
	20	5	0.4	3.42
	30	5.5	0.4	4.05
	20	4	0.4	3.18
	22	5.5	0.7	4.40
	25	6	0.4	3.92
10	12	3.5	1.3	3.80
	12	4	1.3	3.97
	5	3.5	1.1	2.68
	7.5	3.8	1.2	3.25
	9	3	1.1	3.10
	5	3.3	0.8	2.36
	6.8	2.5	0.9	2.48
	9	2.4	0.9	2.69
	8	3.6	0.8	2.85
	5.8	3.7	1.0	2.78
	8	3.7	1.1	3.20
	4.5	2	1.2	2.21
	5.5	3.4	0.9	2.56
	6.5	4	3.4	4.46
	5.5	2.8	1.1	2.58
	4.9	2.9	1.2	2.58
	8.2	3.5	1.4	3.43
	6.6	5.2	1.0	3.44
	6.7	3.2	0.7	2.47
	8.2	1.9	1	2.50
	10	2.4	1.8	3.51
11	30	3.3	0.3	3.10
12	45	10	3	11.08
	20	9	2	7.12
	32	11	7	13.50
	24	7	3	7.96
13	39	12	7	14.89
	37	9	6	12.61
	28	9	6	11.50
	18	8	5	8.95
	15	4	1	3.92
	13	4	1	3.74

* The original diameters (d) were calculated from eq. (7), p. 481. The localities are tabulated under figure 1.

lamellae parallel to the plane of schistosity, and each lamella is moved a certain distance in relation to the underlying lamella. Then the "amount of shear" (Becker, 1893, p. 25) is the distance *divided by* the thickness of the lamellae. The direction of movement is the tectonic *a* axis of Sander (1930), and the *c* axis is perpendicular to the plane of schistosity.

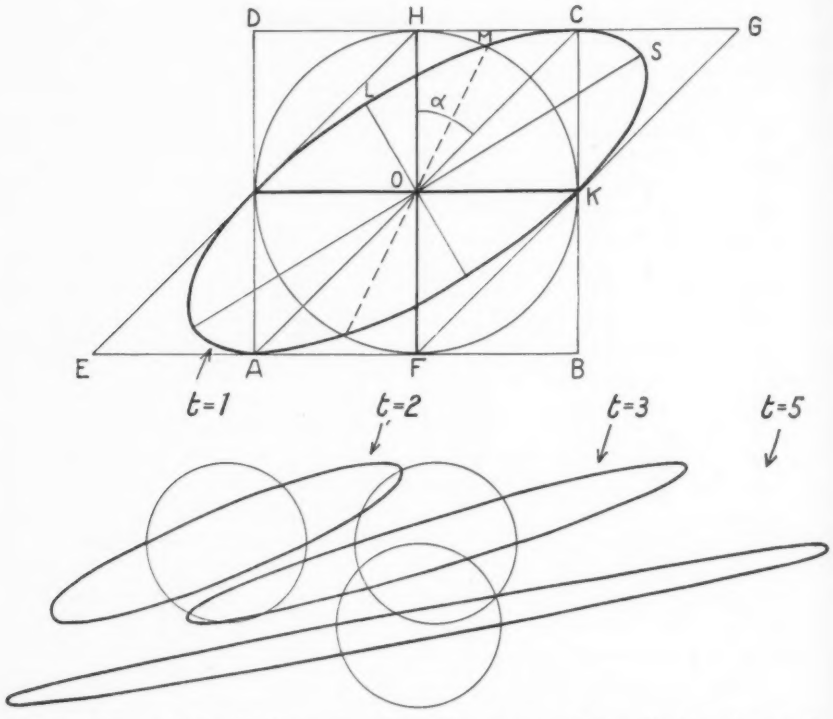

FIG. 2.—Deformation by shearing motion with varying amounts of shear (t), drawn in *ac* sections. *Above:* With $t = 1$ the square *ABCD* is transformed into the parallelogram *EFGH* (*OK* being plane of relative movement), and the circle with radius *OK* is transformed into an ellipse with semiaxes *OS* and *OL*. *Below:* Deformations with amount of shear $t = 2$, 3, and 5.

Figure 2 shows the deformation of a sphere by increasing amounts of shear (t). The sphere with radius *OK* becomes an ellipsoid with semiaxes a = *OS* and c = *OL*, while b (vertical to the paper) remains constant = *OK*. With increasing amount of shear the ellipsoid is increasingly elongated and flattened.

If the radius *OK* = 1, the equation of the circle is

$$x^2 + y^2 = 1 . \qquad (1)$$

By deformation, the circle is altered to a "strain ellipse" (Nadai, 1931, p. 295):

$$x^2 - 2txy + (t^2 + 1)\, y^2 = 1 . \qquad (2)$$

By rotating the co-ordinate system, the formula of the ellipse becomes

$$\frac{x^2}{a^2} + \frac{y^2}{c^2} = 1 \qquad (3)$$

where

$$\frac{a}{c}\Big\} = \sqrt{\tfrac{1}{2}(2+t^2 \pm t\sqrt{t^2+4})} \quad (4)$$

and $a \cdot c = 1$.

The amount of shear (t) is easily defined by $\tan \alpha = t$ (fig. 2), or from equation (4) by:

$$t = \sqrt{a^2+c^2-2}\,. \quad (5)$$

CALCULATION OF THE DEFORMATION

We assume that a spherical pebble is deformed. The volume remains constant:

$$\tfrac{4}{3}\pi r^3 = \tfrac{4}{3}\pi abc\,, \quad (6)$$

$$r = \sqrt[3]{abc} \quad (7)$$

or, when $r = b$,

$$r = \sqrt{ac}\,, \quad \text{or} \quad r^2 = ac\,. \quad (8)$$

Then

$$\frac{a}{r} = \frac{r}{c}\,. \quad (9)$$

This ratio, given as a percentage ($100\ a/r = 100\ r/c$), may be taken as the deformation ratio of the a and c axes.

But it is easy to find that b always deviates more or less from $\sqrt{ac}$. Thus the b axes also undergo a deformation. Then r is defined by $\sqrt[3]{abc}$. Still it is advantageous to use $100\ a/r$ and $100\ r/c$ as the deformation ratios of the axes a and c. Now they differ from each other; and the more they differ, the larger the deformation of b. The latter deformation is given by $100(b - r)/r$, as defined by Cloos (1947, p. 863).

This way of calculating differs slightly from that of Cloos (1947, p. 863). He gives all three components of deformation by the difference from the original radius, in percentage of this radius; for example, for the c axis: $100\ (c - r)/r$. By this procedure Cloos gets various values for the deformation in a and c, even if the deformation is of the ideal type, namely, with no distortion in b. For example, if the dimensions in centimeters are 100, 10, and 1. Then $r = \sqrt[3]{1000} = 10$. Following Cloos, the deformation in a is

$$\frac{100-10}{10} \cdot 100 = 900 \text{ per cent}\,,$$

and in c:

$$\frac{1-10}{10} \cdot 100 = -90 \text{ per cent}\,.$$

By the calculation suggested above the corresponding properties are:

$$\frac{100}{10} \cdot 100 = 1{,}000 \text{ per cent}\,,$$

$$\frac{10}{1} \cdot 100 = 1{,}000 \text{ per cent}\,.$$

This property, 1,000 per cent, expresses the deformation of the rock.

In figure 3 the deformation of a is plotted against the corresponding deformation of c. Points on the broken line have equal deformation ratios of a and c, and therefore they mark the pebbles which show no distortion of b. Most of the points lie above the line, indicating a higher deformation of c than of a. This means that b has been extended. Another expression for this relation is obtained by a plot analogous with figure 12 of Cloos (1947, p. 889), where the deformation of b is plotted against the deformation of a (fig. 4).

Figure 4 shows that the pebbles under consideration exhibit a much greater deformation in b than did the oöids of Cloos. In addition to an extension in b, reduction in b is not rare. As already seen from figure 3, deformation in a (and c) is also, in general, much more intense than that observed by Cloos.

There is another way of characterizing

the deformation, defined by the amount of shear. For example, in the above-mentioned case, the deformation is 1,000 per cent, but the amount of shear is about 10, according to equation (5). The property 10 means that, if the conglomeratic bed, 1 meter thick, is deformed homogeneously, the upper part of it has moved 10 meters in relation to the lower part. Thus the amount of shear gives a much clearer picture of the degree of deformation than does the deformation percentage, and it is called the "gradient of deformation" in the following discussion.

An ellipsoidal pebble with semiaxes a, b, and c may be characterized by the double ratio a : b : c. In the form

$$\frac{a}{c} : \frac{b}{c} : 1$$

the ratios are independent of the absolute size, and each pebble may have its

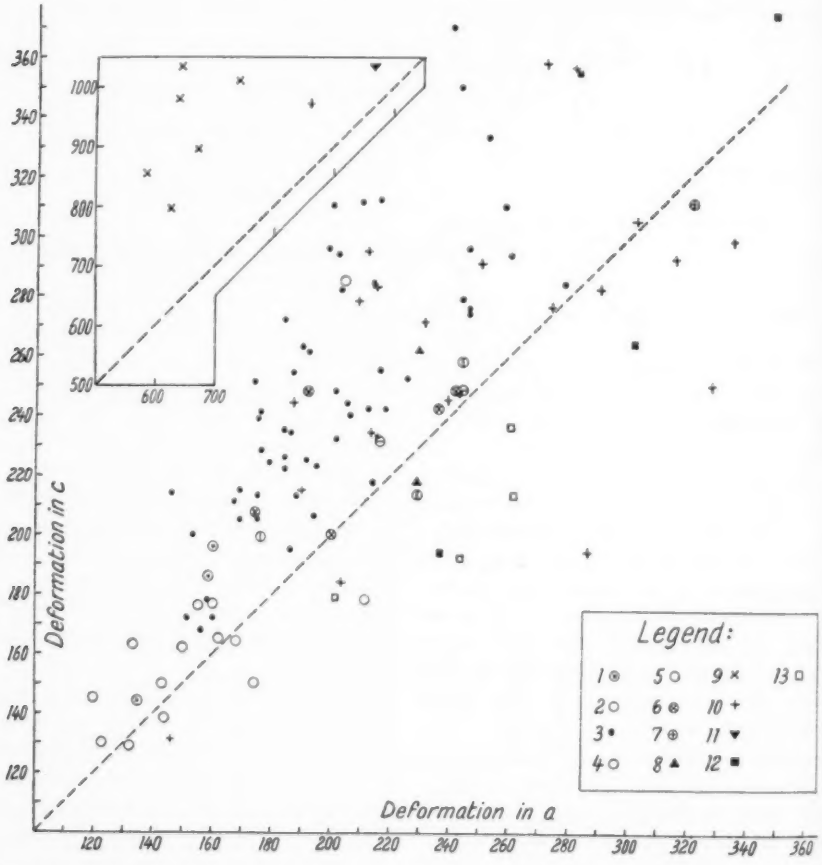

FIG. 3.—Deformation of a axes in relation to deformation of c axes, given by the expression 100 a/r and 100 r/c, viz. The numbers in the legend refer to the localities, as tabulated under fig. 1. *Inserted:* Highly deformed conglomerates.

deformation marked by the values a/c and b/c. Equation (4) is deduced from the deformation of a unit sphere (r = 1). Then b = r = 1, and the axial ratios are a : 1 : c or

$$\frac{a}{c} : \frac{1}{c} : 1 .$$

By giving t the values 1, 2, 3, etc., in equation (4), the values a/c and 1/c may be given as a function of t. These values give the two full-line curves of figure 5, one for a/c and one for b/c. In this diagram the deformed pebble may be easily plotted as a point by the following procedure: From a = 4, b = 2, and c = 1, the values a/c = 4 and b/c = 2 are calculated. The point on the curve for b/c with an ordinate equal to 2 is found, and vertically above it, with an ordinate equal to 4, the point is plotted. This point will lie on the curve for a/c, if the pebble is deformed by a pure shearing motion. Thus figure 5 indicates in a new way that the deformation of most of the measured pebbles deviates from pure shearing motion. But the plots, which fall near the a/c curve, directly define a gradient of deformation.

PROBLEMS OF THE DEFORMATION

The plots of figures 3, 4, and 5 show that the deformation of the pebbles was not solely a shearing motion. The movement, therefore, may be of a more complicated type, and, in addition, some other factors may influence the result. Various facts suggest such assumptions.

First, the present orientation of the pebbles does not agree with the theory of shearing motion, because then the long axis of the ellipsoidal pebble should form a certain angle with the plane of schistosity. On figure 2 this angle is (90 − a)°. In mildly deformed conglomerates the angle should be easily observed. However, the long axis is lying in the plane of schistosity, or, in highly deformed conglomerates, forms a small angle with the plane in certain localities. The fact that the long axis, a, mostly

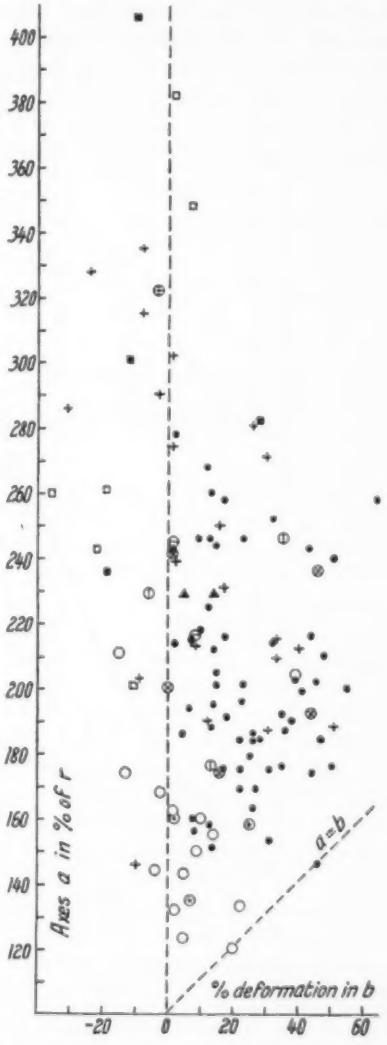

FIG. 4.—Deformation of b axes (given by 100 [b − r]/r) in relation to deformation of a axes, given 100 a/r as in fig. 3.

coincides with the tectonic axis *a*, may be explained either as being caused by an external rotation or by another type of deformation than shear on the plane of schistosity. I consider the second possibility as improbable, and I suggest that the whole pebble has been rotated, simultaneously with the shear deformation.

By inhomogeneous strain the pebbles should get a shape deviating from that of an ellipsoid. In *ac* and *bc* sections the pebbles are observed to have outlines which are scarcely exact ellipses. Even a flat ellipse is rounded at the ends, but the pebbles are commonly sharp. It is, however, difficult to get a quantitative

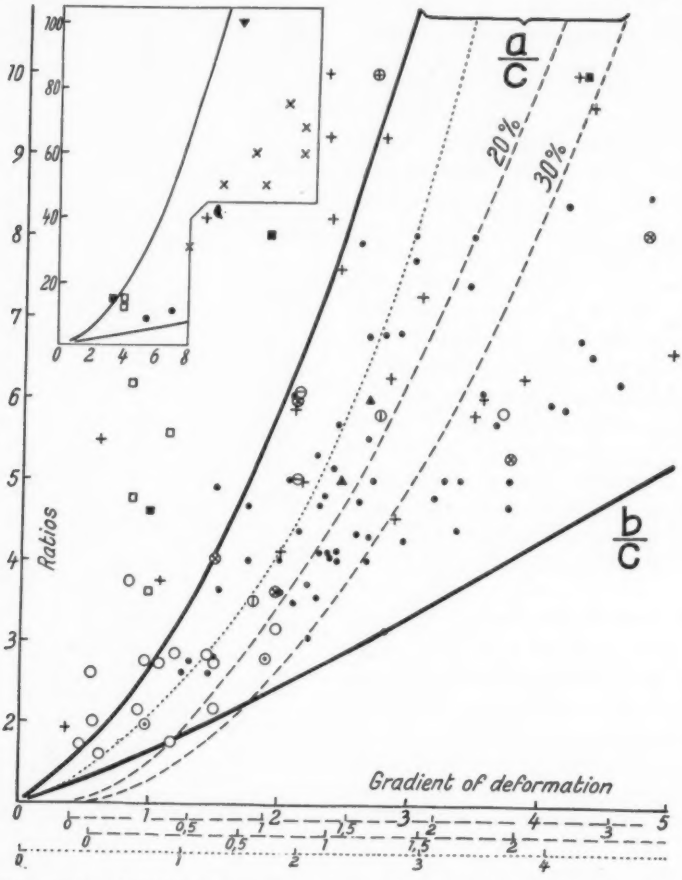

FIG. 5.—Deformation of pebbles by plots of the axial ratios a/c : b/c : 1 in relation to the gradient of deformation. The full-line curves give the values of a/c and b/c for a varying gradient by shearing motion. The values are calculated from equation (4). A pebble with axes a, b, and c is plotted as a point by the ratios a/c : b/c : 1. The value b/c is found on the curve for b/c, then the point has the same abscissa, and the ordinate a/c. If the point falls near the full-line curve for a/c, the deformation is a shearing motion, and the gradient of deformation may be read. Curves for a/c for shearing motion combined with elongation in b (*broken lines*) and for nonspherical shape of the original pebble (*dotted line*) has been constructed (see text, pp. 485 and 486). *Inserted:* Plots of highly deformed pebbles. The same designations are used. Only the shearing-motion curves for b/c and a/c are drawn.

measure of these deviations. The two major problems of the deformation are alteration of the b axes and the primary shape of the pebbles. These problems are discussed at some length below. They are considered independently.

DISTORTION OF THE *b* AXES

If we assume that the deformed pebbles originally had a spherical shape, the present shape should give a true account of the deformation process. Accordingly, the shape should give quantitative expressions for the deformation in *b*.

Cloos pays considerable attention to the "distortion parallel fold axes" (pp. 886–892). He specially emphasizes the importance of elongation in *b* in arcuate orogens.

The Caledonian orogen forms an arc just in the area from which the Sparagmites have been thrust. Regional maps show that the orogen alters its direction a certain angle in the course of 300 km. The angle may be estimated to be 10°–20°. The exact elongation may then be calculated, if the necessary properties are determined. Let us assume that the arcuation is 10° over a width of 300 km.

In order to calculate the pebble deformation in a Sparagmite nappe with dimensions 300 × 300 km., some further simplifications must be made. If the pebbles have an external rotation simultaneously with the shearing motion, the calculation of the corresponding strain ellipse becomes very intricate. Therefore, the rotation is considered to occur after the shearing motion. Then the b axes are elongated, and the c axes are reduced, while the a axes remain unchanged.

In the front of the thrust masses the increase of the nappe parallel to the b axes is sin 10° · 300 km. = 52 km., or nearly a 20 per cent increase. A corresponding reduction occurs to the c axes.

From figure 4 the deformation in *b* may be read directly. But in order to express deformation in *a* and *c* by the easily understandable "gradient of deformation," the deformation must be considered as shearing motion combined with irrotational homogeneous strain parallel to the b axis.

A pebble with a radius r is deformed by shearing motion to an ellipsoid with semiaxes a, b = r, and c. The elongation in b of, say, 20 per cent changes the dimensions to a, 1.2 b, and c/1.2. Instead of

$$\frac{a}{c} : \frac{b}{c} : 1 ,$$

the ratios are

$$1.2\frac{a}{c} : (1.2)^2\frac{b}{c} : 1 .$$

From the latter ratios a number of values for the varying gradient *t* may be calculated (eq. [4]). These values define two curves.

There are two ways of drawing these curves for comparison with the ordinary shearing motion. Either the two curves may be drawn in figure 5 with the same unit of gradient as the shearing motion, or the already existing curve for b/c may be used as a reference curve, as explained for the plotting of measurements. For a certain *t* the ordinate $(1.2)^2(b/c)$ is found on the curve for b/c, and vertically above this point the ordinate $1.2(a/b)$ is plotted. Points for various *t* values define a curve. By this procedure the deformation with elongation in b is demonstrated by a new curve, which has other units for the gradient *t* on its abscissa axis.

The latter way of establishing curves is followed. On the new curves (broken curves on fig. 5) we should find all the points, which represent pebbles deformed by shearing motion with elongation in *b* equal to 20 per cent and 30 per cent, and a varying gradient of deformation. By extrapolation the extension in *b* and the

gradient of every observation may be read from figure 5.

THE ORIGINAL PEBBLE SHAPE

No conglomerates have pebbles with a shape which mathematically is a sphere. Undeformed quartz conglomerates, similar to the described ones, are found along the southeastern border of the Sparagmite formation. The pebbles of these conglomerates have a shape which varies to a certain degree. All the pebbles are well rounded, and most of them are spherical or nearly so, but some pebbles are ellipsoidal, with one long axis in the bedding plane, or they are flat in this plane, that is, they are rotational ellipsoids with the rotation axis (short axis) vertical to the bedding plane. This latter shape is considered as the most typical of deviations from a spherical shape. The author has tried, therefore, to calculate the deformation of such an ellipsoid.

If the ellipsoidal pebble has the semi-axes a', $a' = b'$, and c', its equation is

$$\frac{x^2}{a'^2}+\frac{y^2}{a'^2}+\frac{z^2}{c'^2}=1\ ,\ \text{where } a'>c'\ . \quad (10)$$

In the ac section the corresponding ellipse,

$$\frac{x^2}{a'^2}+\frac{z^2}{c'^2}=1\ , \quad (11)$$

is transformed by shearing motion into another ellipse:

$$c'^2x^2-2c'^2txz+(c'^2t+a'^2)\,z^2=a'^2c'^2\ . \quad (12)$$

By rotation of the co-ordinate system, the formula of the ellipse becomes

$$\frac{x^2}{a^2}+\frac{z^2}{c^2}=1\ , \quad (13)$$

where

$$\left.\frac{a}{c}\right\} = \sqrt{\tfrac{1}{2}\left(a'^2+c'^2(t^2+1)\ \pm\ \sqrt{(a'^2+c'^2[t^2-1])^2+4t^2c'^4}\right)}\ . \quad (14)$$

We choose $a' = b' = 1$, $c' = \frac{3}{4}$, as probable values for an ellipsoidal pebble. This shape is one of the most extreme deviations from the sphere. The deformed pebble has the axial ratios ($b' = 1$ is constant) $a : 1 : c$. On the usual form:

$$\frac{a}{c}:\frac{1}{c}:1\ .$$

By giving t the values 1, 2, 3, etc., we obtain a set of values for a/c and for 1/c which define two curves. In order to compare the plotted observations of figure 5 with this new case, the two curves are substituted by one new curve in figure 5, using the old b/c curve as a reference curve, as described for the curves showing the extension in b.

The pebbles which have their points lying on the dotted curve of figure 5 may accordingly originate from deformation by shearing motion, if the original pebble was a flat rotational ellipsoid with semi-axes 1, 1, and $\frac{3}{4}$. And, with the original pebble shape lying between this ellipsoid and the sphere, the points of the deformed pebbles should lie statistically between the full line and the dotted a/c curve, if the deformation type was shearing motion. However, relatively few points lie between the mentioned curves, and this fact proves that the deviation from the spherical shape of the original pebbles is not a factor of primary importance in relation to extension in *b*.

CONCLUSIONS

The facts and calculations presented above show that the deformation of the measured pebbles is of a very complex nature. The conclusions cannot be as quantitative as intended, and they may

be subject to revision when more material is collected. But some qualitative results are obtained, and some semiquantitative estimates are given.

1. Shearing motion is considered as the main element of the deformation, but figure 5 indicates that, in addition, a deformation in b is evident. The deformation varies from increase (up to 30 per cent) to decrease (to about 30 per cent). The deviation from a sphere in the original pebble shape does not influence severely the significance of the figures but probably causes the variation within each locality.

2. The deformation in a and c increases from the border of the orogen toward the central zone. For the different localities the gradient of deformation may be read as follows: Nos. 1 and 2: gradient below 1. No. 3: mostly between 1 and 1.5; a few above 2. Nos. 4, 5, 6, 7, and 8: between 1 and 2. No. 10: mostly between 2 and 3. No. 9: probably from 6 to 10. No. 11: 10. Nos. 12 and 13: between 2 and 4.

3. The mean extension in b may be estimated as follows for the different localities: Nos. 1 and 2: uncertain. No. 3: 20 per cent. Nos. 4, 5, 6, 7, and 8: between 10 and 20 per cent. No. 10: 10 per cent. No. 9: between 20 and 30 per cent. No. 11: 0 per cent. Nos. 12 and 13: mostly a pronounced reduction in b (20 per cent).

Thus the estimates show that the extension in b decreases from the border toward the central zone. In highly deformed conglomerates even a pronounced reduction occurs.

The decrease of the extension may be connected with the arcuated orogen.

ACKNOWLEDGMENTS.—The field observations for the present paper were collected during geological mapping for the Geological Survey of Norway. The author is indebted to Professors F. J. Pettijohn and R. B. Balk for corrections and suggestions.

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GEOLOGICAL NOTES

A STEREOGRAPHIC CALCULATOR[1]

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DESCRIPTION OF INSTRUMENT

The stereographic calculator here described represents essentially a combination in one physical unit of the stereographic net, Penfield paper, and a stereographic scale, thus eliminating certain mechanical difficulties in manipulating the units separately. The combination results in a sturdy, lightweight instrument, like a slide rule in action, with which problems solvable by the stereographic projection can be done easily and rapidly either in the field or in the office.

The instrument (fig. 1) is composed of three main parts: (1) a stereographic net on a circular disk of metal or cardboard; (2) a graduated circle on a circular disk of transparent plastic, the front of which is frosted to take pencil; (3) a circular metal housing in which the disks carrying the stereographic net and graduated circle can revolve freely. This metal housing is made in two parts, a back and a front, so that it may be opened and the disks inserted. A circular portion of the back half is cut away so that the disk with the stereographic net can be revolved with the fingers. The front half is cut away so as to expose the net and graduated circle in a semicircular opening and produce a straight edge which bisects the stereographic net. As much of the remainder of the front can be cut away as will produce better visiblity of the net and yet retain the strength of the straight edge as well as of the metal housing itself. On the straight-edge portion of the front a stereographic scale is inscribed.

In the instrument prepared as a model, the housing was made of magnesium, although that metal proved somewhat unsatisfactory, in that it marked the frosted disk relatively easily. A stereographic net, 10 cm. in diameter, on cardboard backing, prepared by Fisher and distributed by Ward's Natural Science Establishment, Inc., of Rochester, New York, was used and cut into disk form to make the rear disk. Ribs were glued to the back of the disk, so that torque could be applied to rotate the disk easily. The front disk was made of ordinary transparent celluloid; the graduated circle was inscribed by hand with a stylus, and the front was frosted by a few minutes' grinding with fine-grained abrasive on a glass surface.

USE OF INSTRUMENT

Many simple problems, such as calculating the apparent dip of a bed in any vertical section cutting the bed (fig. 2) can be solved merely by rotating the two disks and scale into the proper positions with respect to each other and reading the answer directly. Other problems require marking a point or line on the frosted surface of the disk containing the graduated circle and then rotating the disks and scale to a position where the answer may be read. The method for solving structural problems as described by Bucher (1944) will suggest clearly the procedure for using the present instrument.

Although any problem involving stereographic projection can be done on this instrument, the solution of certain problems can be further simplified by using disks with different combinations of nets inscribed on them. For example, if one has numerous problems involving the determination of the trend and plunge of the line of intersection of two planes, a combination of two disks, each inscribed with nothing but great-circle

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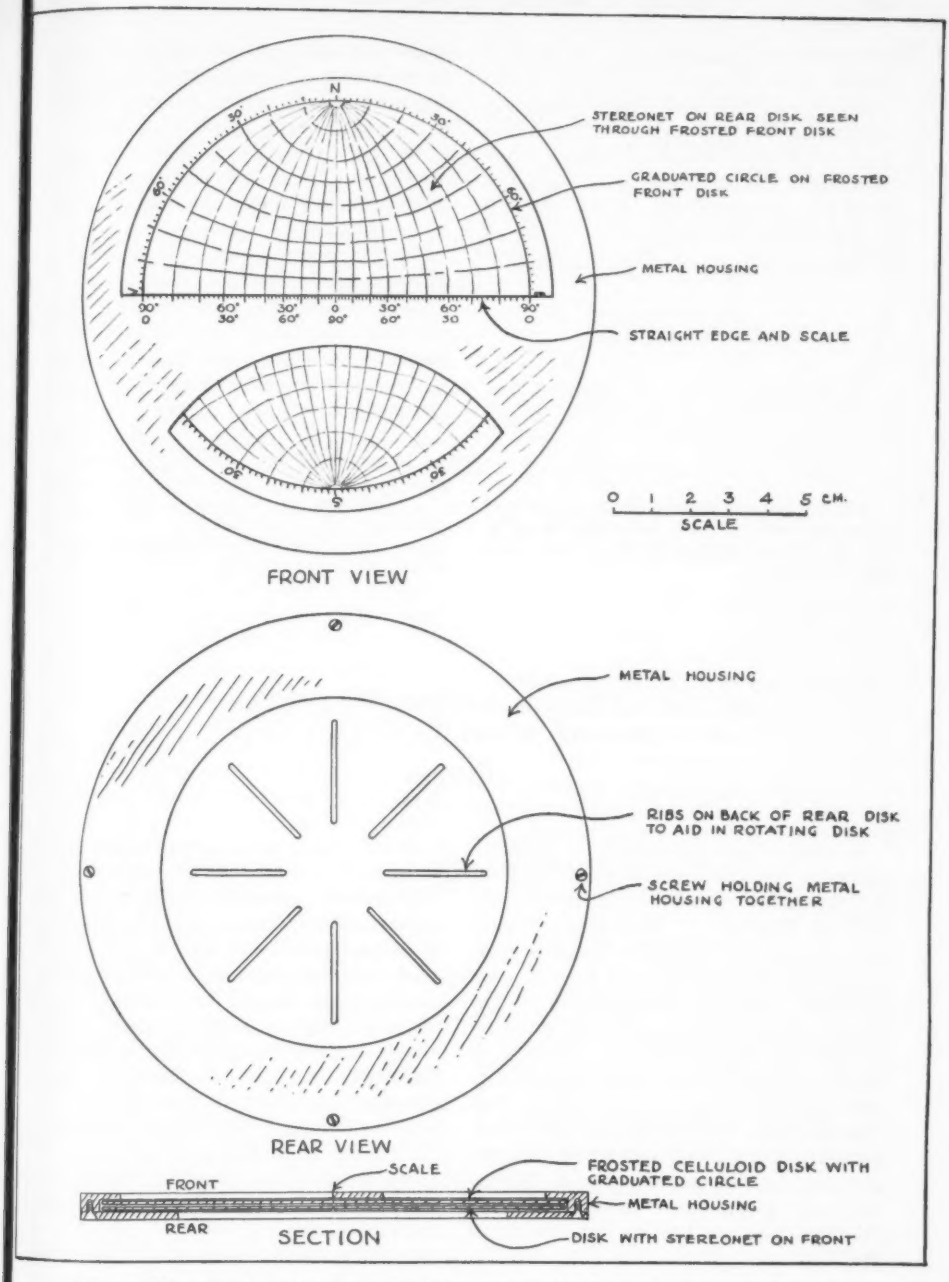

FIG. 1.—Diagrammatic sketch of stereographic calculator, showing construction of instrument

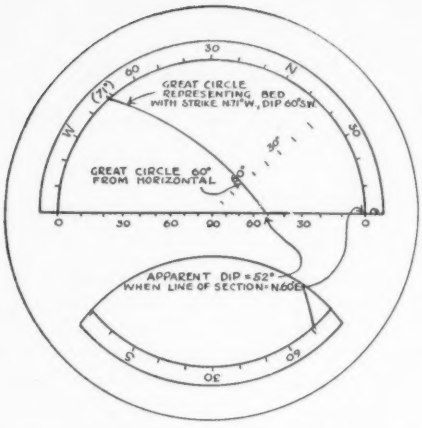
GREAT CIRCLE REPRESENTING BED WITH STRIKE N71°W, DIP 60°SW
GREAT CIRCLE 60° FROM HORIZONTAL
APPARENT DIP = 52°
WHEN LINE OF SECTION = N60E

FIG. 2.—Example showing direct solution of apparent dip problem, where bed strikes N. 71° W., dips 60° SW., and strike of the line of vertical section is N. 60° E.

projections and degree divisions around the circumference of the projection, makes solution a matter of reading answers directly after proper rotation of the disks. Inasmuch as the metal housing can be taken apart readily, any of a variety of disks can be substituted to simplify the solution of certain problems.

The instrument has proved very satisfactory as a sturdy field and office tool for calculation of structural problems. In the field the easy solution of problems makes it possible to realize relations that might otherwise be overlooked, and pertinent study is suggested.

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DETERMINATION OF SODIUM AND POTASSIUM IN SILICATE MINERALS AND ROCKS[1]

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Several authors have described improvements of the Lawrence-Smith method for the determination of alkalies in silicates. Willard, Liggett, and Diehl (1942, p. 234) and Marvin and Woolaver (1945, p. 554) recommend hydrofluoric acid together with perchloric acid for disintegration of the sample. This solvent is also suggested by Lundell and Knowles (1927, p. 849). To remove excess hydrofluoric acid, Willard, Liggett, and Diehl (1942, p. 234) distil it off as siliconfluoride, whereas Marvin and Woolaver (1945, p. 554) fume it off by repeated evaporations with perchloric acid. To convert the perchlorates into oxides and chlorides, they use heating in a furnace at 550° C.

In both cases special precautions have to be taken to get rid of interfering elements. If phosphorus is present, the methods have to be modified. The procedures appear to require more skill and training and to take more time than the Lawrence-Smith method. The only step in the Lawrence-Smith procedure that needs great caution is the mixing of the sample with the flux and the transferring of it to the crucible. The mixing, however, can be done without any perceptible loss after a little training. The great advantage of the fluxing procedure is that it leaves the alkalies in solution, free from interfering elements, after but a few, simple operations. Small amounts of calcium and magnesium might contaminate the alkali chlorides and will effect the determination

[1] Manuscript received February 11, 1948.

of sodium if this is done by deducting it from the sum of the chlorides. By determining the sodium as $NaZn(UO_2)_3(CH_3COO)_9 \cdot 6H_2O$, however, this uncertainty can be eliminated.

Miller and Traves (1936, p. 1390) give a thorough discussion of the determination of sodium as the triple salt and give some experimental facts about the influence of calcium. Their results show that small amounts of calcium are of little or no consequence. After adding 1 gm. of $CaCl_2 \cdot 6H_2O$ to a solution from which 14.4 mg. of sodium chloride is to be precipitated, a double precipitation gives results only 0.1–0.2 per cent too high. (First precipitation was approximately 2 per cent too high.) The amount of calcium by the second precipitation will be of the same order of magnitude as in the case of the alkali chlorides after the Lawrence-Smith procedure has been followed.

During a short stay at the Department of Geology of the University of Chicago, I had the opportunity of making a few experiments investigating these problems.

The alkali content of several mineral and rock specimens was determined by the Lawrence-Smith method, and the potassium was determined as K_2PtCl_6. Instead of discarding the filtrate after the precipitation, the sodium was determined as the triple salt after the only interfering element, platinum, had been reduced by evaporation of the alcoholic filtrate in a conical flask on the waterbath. If, after evaporation to dryness, the reduction is not complete, a few milliliters of alcohol are added and evaporated as before. The reduction is complete when the yellow color of the platinum complex has disappeared. The reduced platinum occurs either as a powder or in thin flakes and is easily washed free of chlorine ions. Sodium was precipitated from the filtrate as the triple salt, by the procedure described by Kolthoff and Sandell (1945, p. 416).

B. Bruun and R. Higazy in the laboratory of the Geology Department made several determinations of sodium and potassium in rocks and minerals, with the Lawrence-Smith procedure as described by Washington (1930, p. 222). From the filtrates after the precipitation of potassium as chloroplatinate, the sodium was precipitated as triple salt. The results are given in table 1:

TABLE 1

PER CENT Na_2O

Higazy's Determination	Triple Salt Determination
1.32	1.21
11.03	10.78
2.98	2.98

Bruun's Determination	Triple Salt Determination
2.99	3.00
10.35	10.07
1.75	1.74
2.12	2.03

There is a tendency to get somewhat lower results by determination as triple salt. This must be ascribed to the fact that calcium and magnesium are not always completely removed before the sum of the alkali chlorides is determined, but this does not effect the determination of sodium when it is precipitated as triple salt.

This procedure may be especially useful when the amount of sodium is small compared with that of potassium.

ACKNOWLEDGMENTS.—I take pleasure in expressing my thanks to Dr. W. H. Newhouse, chairman of the Department of Geology at the University of Chicago, who placed all facilities of the department at my disposal and made this investigation possible. My thanks go to Dr. T. F. W. Barth, of the same department, for his never failing interest and encouragement.

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SHALLOW-WATER ORIGIN OF RADIOLARITES IN SOUTHERN TURKEY[1]

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Radiolarites in Turkey and also in many other parts of the world are closely associated with basic volcanic rocks and manganese concretions and are shallow-water deposits connected with volcanic activity either continental or submarine. Fossil, abyssal radiolarite deposits are probably exceptional occurrences.

In the Malm (Upper Jurassic) of southern Turkey, particularly in the central Toros geanticline (Tromp, 1947, p. 367, Table II), a thick series of sediments occurs which has been described by Ortynski as "red beds" and by Lahn, Arni, Blumenthal, and others as "radiolarite series." It is composed mainly of red shales with many red and a few green chert beds, whose number increases upward in this section. Several of the red chert beds are rich in radiolarians, a fact which gave rise to the name "radiolarite series." Many of the chert beds are nonfossiliferous, however, and the name "red beds" would be preferable. But the name "radiolarites" is so generally used for all these red chert beds, both in Turkey and in other parts of the world, that we shall continue to use this latter name.

Most geologists consider the radiolarites an indication of deep-sea environment, as similar deposits are known in present-day oceans at depths of over 4,000 meters. However, most, if not all, radiolarite deposits in Turkey are shallow-water sediments. This can be proved as follows:

1. Red chert beds, with or without radiolarians, are a most common feature not only in the Jurassic radiolarite series but also at the border of large serpentine massives (mostly gabbroic rocks, metamorphosed later into serpentines). They are covered by basal conglomerates or other coarse sediments belonging to the basal portion of a younger overlapping series, indicating a shallow-water origin of the radiolarites. If those chert beds had been formed in an older period and did not belong to the new cycle of sedimentation, it would be remarkable that the erosion always preserved both the serpentines and their red cherty border zones.

2. The Jurassic series in southeastern Turkey is underlain by the so-called "flysch series," which is composed mainly of nonfossiliferous sandstones, many of which are very coarse.

3. At the end of the period of the radiolarite series, the central Toros geanticline rose above sea level, resulting in the Lower Cretaceous unconformity. This suggests shallow-water conditions at the end of the radiolarite period.

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There is also another phenomenon which supports the assumption that radiolarites in Turkey are shallow-water deposits, explaining, at the same time, the origin of these radiolarites.

The radiolarites and red cherts in southern Turkey are always associated, either directly or indirectly, with basic intrusions or extrusions. First, at the end of, and perhaps even during, the Upper Jurassic, basic intrusions took place. Second, during the beginning of the deposition of the Upper Jurassic sediments, several of the Paleozoic serpentine massives were above sea level, mainly because of tilted block movements at the end of the Dogger. Because of erosion, much of the flysch series is missing, and the radiolarite series directly overlies the old serpentine cores (for example, near Egridir).

The influence of basic igneous rocks on radiolarite development may be explained as follows: If basic rocks are dissolved in the sea or if volcanic submarine exhalations and extrusions took place, the sea water would become very rich in silicic acid, which makes the milieu very favorable for the growth of radiolarians. This assumption is supported by various observations in other countries.

1. Recent plankton researches of Johnstone and others showed that, if the silicic acid content of sea water increases, the growth of siliceous algae (diatoms) is promoted. This explains, for example, the occurrence of thin beds of diatomaceous earth, alternating with tuff deposits, near Cheribon, south of Darma, Java, Dutch East Indies.

2. L. van Houten (1930) described radiolarite deposits, alternating with basic volcanic rocks, from the "Buchensteiner Schichten" (Ladinian) in the Dolomites. Next to these deposits large coral limestone reefs occur, indicating a shallow-water origin.

3. Dewey and Flett described Devonian radiolarite deposits from Cornwall, which alternate with lavas and shallow-water deposits.

4. Carstens described radiolarites from the Bymarck formation (Silurian) of the Trondhjemer syncline (Norway), which occur together with basic volcanic rocks of a shallow-water series.

5. J. B. Scrivenor (1912) described a Permo-Carboniferous section in Malaya which is called the "Pahang volcanic series." It is characterized by basic volcanic rocks (diabase tuffs, porphyries, porphyry tuffs, lahars, etc.) with clay slate intercalations and contains fossil plant remains and many radiolarites.

G. A. F. Molengraaff (1909) described similar formations from central Borneo, Dutch East Indies, as the "Danau formation." It is not so rich in plant remains but is very rich in radiolarites and manganese concretions (Molengraaff, 1915). The radiolarites alternate with coarse clastic sediments. J. Wanner (1921) and others, who believed in a deep-sea origin, supposed that these clastic sediments were deposited in small but very deep basins. But as shallow-water reef limestones also occur near by, this explanation seems erroneous.

The occurrence of many manganese concretions together with radiolarites is very characteristic of the Recent deep-sea radiolarian ooze, and this association is therefore often used as an argument for the deep-sea origin of sediments, for example, of the above-mentioned Danau formation of Borneo. However, the association of basic volcanic rocks and sedimentary manganese ore deposits is common in Turkey, for example. Such deposits were recently described by P. de Wijkerslooth (1943).

In northern Anatolia, in the so-called "Pontic geanticline" (Tromp, 1947, p. 368), most manganese ore deposits occur in the Upper Cretaceous flysch deposits, which are composed of an alternation of marls, andesites, and andesitic tuffs, in many cases rich in "jaspis" beds. According to Wijkerslooth, the original manganese deposition is due to volcanic exhalations. The manganese ore was later dissolved and redeposited in the form of concretions.

In the area between Ankara and Karabük

(central Anatolia), several small manganese deposits occur together with Jurassic radiolarites and serpentine massives, etc. The same association has been observed in the central Toros geanticline.

In the western part of the central Anatolian Plateau (Tromp, 1947, p. 368), the Upper Paleozoic beds are composed of limestones, graywackes, and radiolarites with many basic effusive rocks (diabase, etc.). Manganese interlayers are also quite common in these beds.

All these different phenomena sufficiently indicate that the association of volcanic rocks, radiolarian chert, and manganese concretions is a normal one, which might be due to deep-sea deposition but which, in general, is typical of shallow-water deposition connected with volcanic activity, either continental or submarine.

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REVIEWS

Outlines of the Geography, Life, and Customs of Newfoundland-Labrador (the Eastern Part of the Labrador Peninsula). By V. TANNER. 2 vols. Cambridge: Cambridge University Press, 1947. Pp. 906; figs. 342.

In these two volumes Professor Tanner, of the University of Helsingfors, summarizes available information on the geography of eastern Labrador. Much original material, based on observations made during the Finland-Labrador Expedition in 1937 and the Tanner Labrador Expedition in 1939, is presented. Volume I is divided into five parts, which deal with the geology, oceanography, meteorology and climate, and plant and animal life; Volume II is given over entirely to human geography.

Geologists will be interested mainly in the first 254 pages describing the principal geologic features: the pre-Cambrian rocks and their structure, the Tertiary peneplain, effects of glaciation, and raised shorelines.

The summary of pre-Cambrian geology is based on studies by E. H. Kranck, a member of the 1937 expedition, and upon earlier studies by Daly, Coleman, Odell, and others. The following units are described in considerable detail:

Eruptive rocks of undetermined age
- Strawberry granite and related granites(?) and syenites

Algonkian series
- Double Mer and Lake Melville beds (Keweenawan)
- Ramah and Kaumajet (Cape Mugford) beds (Huronian)

Archean series
- "Domino gneiss"
- Makkobik granite
- Anorthosite gabbro
- Migmatite gneiss with mafic or lime-rich inclusions
- Migmatitic, microcline-rich gneisses and granite gneiss
- Aillik quartzite, conglomerate, and other sediments

Two major orogenic belts are recognized: (1) the Labrador mountain chain adjoining and paralleling the present coast, which was formed at the close of the Archaeozoic, and (2) the Labrador Range in the interior, which trends north-northwest from Mishikaman Lake to the west shore of Ungava Bay and was formed in late Huronian time.

The Cenozoic record, in which Tanner was particularly interested, begins with uplift of the "general peneplain surface" in the Pliocene. This surface is considered to be *metachronous* in the sense that it is composed of several facets developed at various times since the Huronian orogeny. The major preglacial valleys were initiated by uplift of the peneplain, and many persist to the present time. During the Pleistocene, Wisconsin ice from the west is believed to have covered the entire region, including the summits of the Torngak Mountains, which Daly, Coleman, and others previously had considered nunataks. No evidence of earlier glaciation is reported. Upwarp of the region following glaciation is evidenced by abundant raised shore features at various elevations. Because of uncertainties, only general interpretations are made from the marine limits, and there is no attempt to apply the epeirogenic spectrum graphs developed by the author in Fennoscandia.

As is clearly recognized, much of the geology of Labrador remains for future investigation. Professor Tanner's study, however, is a valuable summary and is of special significance because of the author's long experience with strikingly similar problems in Fennoscandia.

L. H.

Aids to Geographical Research. By JOHN KIRTLAND WRIGHT and the late ELIZABETH T. PLATT. ("American Geographical Society Research Series," No. 22.) 2d ed. New York: Columbia University Press, 1947. Pp. 331. $4.50.

The first edition of this useful reference work was published in 1923 as No. 10 of the "American Geographical Society Research Series" and has long been out of print. The new edition thus is completely revised and enlarged. It is primarily a guide to geographical bibliographies, periodicals, atlases, and gazetteers; but it also includes considerable geological reference mate-

rial, especially in the fields of physical geography and related sciences. The book is divided into three parts: Part I, "General Aids"; Part II, "Topical Aids"; and Part III, "Regional Aids and General Geographical Periodicals."

L. H.

Correlation of Pleistocene Deposits of Nebraska. By G. E. CONDRA, E. C. REED, and E. D. GORDON. (Nebraska Geol. Survey Bull. 15 [1947].) Pp. 73; figs. 15.

Because of the unusually complete stratigraphic record of the Pleistocene found in Nebraska, this résumé is of general interest. The results of recent subsurface studies and soil investigations are emphasized, and a revised Pleistocene classification is presented. Two new stratigraphic names are proposed: (1) the *Seward formation*, as an eastern facies of the Ogallala and (2) the *Crete sand*, formerly described by Lugn as "valley phase" of the Loveland loess, as a channel deposit of Illinoian age. The loess deposits (Loveland, Peorian, and Bignell) are believed to be mainly eolian in origin, derived principally from silty Tertiary formations of the High Plains, and to be of interglacial rather than glacial age. Brief concluding sections deal with Pleistocene terraces, water-table fluctuations, economic relations, and plants and animals. Many Pleistocene geologists will not agree completely with the interpretations in the report, but most will find them stimulating and helpful.

L. H.

1945 Reference Report on Certain Oil and Gas Fields of North Louisiana, South Arkansas, Mississippi and Alabama. (Shreveport Geol. Soc., vols. 1, 2.) Shreveport, 1947 (planographed). Vol. 1, pp. xxi+328; pls. 3. $12.50. Vol. 2, pp. xii+329–503, pls. 4–12. $10.00.

The greater portion of these two volumes is a summary, in telegram style, of the geologically important facts of 25 fields in Arkansas, 48 in Louisiana, 16 in Mississippi, and 1 in Alabama —a total of 90 fields. The data are grouped under the following subjects: location, prediscovery data, discovery well, well data, structure, producing zones, well spacing, allowable, development through January 1, 1946, deepest well in field, proven area, production data through January 1, 1946, pipe-line data, and miscellaneous. These data are tabulated on twelve pages following page 503, volume 2. Structure contour maps, cross sections, well logs, and other diagrams accompany the descriptions.

To students of Gulf Coast stratigraphy, two chapters by Hazzard, Blanpied, and Spooner (pp. 472–503 of vol. 2) on Cretaceous correlations and on the formations that underlie the Smackover limestone in southern Arkansas will be particularly interesting. New information from deep wells indicates that certain modifications of the earlier correlations will be necessary. Rocks of probable Permian age are believed to have been reached in several deep wells. Numerous correlation charts and diagrams illustrate the discussion.

The compilation of this immense amount of material has been difficult, as explained in the Preface, and several pages of errata had to be added. Interspersed with the text are advertisements of thirty-three firms. The typography of the books is good, but a few pages are missing, and some folded plates have been clipped in so that they must be cut to be used. The high cost of these volumes is most regrettable.

R. B.

Papers from the Geological Department, Glasgow University. (Glasgow University Publications, vols. 69–71; Geological Department Publications, vols. 20–22.) Glasgow: Jackson, Son & Co., 1947.

The first two volumes consist of twenty-seven reprints of octavo size originally published during the period 1937–1945. The third volume consists of twelve papers of larger page size, printed in 1937–1945 and now gathered into a quarto volume. The thirty-nine articles form a cross section (not a complete set) of the publications of the University's Geological Department and include a great range both of subject and of importance. Papers deal with such diverse branches of geology as paleontology and evolution (fifteen papers), structural geology (six papers), petrology (four papers), and smaller numbers concerned with geomorphology, economic geology, and techniques. They range in length and value from a page of strictly local interest to 65 pages with good discussions of principles, descriptions of new fossil species, or original rock analyses. Perhaps the greatest value of these volumes is that they bring to the attention of a larger circle of geologists some useful

papers originally printed in what are, to most American geologists at least, seldom seen publications.

M. S. C.

Transactions of the Geological Society of South Africa, vol. 47 (January to December, 1944), Johannesburg: Hortors Ltd., 1945. 2G.

This issue of the *Transactions* contains the following papers: (1) "Hillslopes and Dongas," by L. C. King and T. J. D. Fair; (2) "Fossils from the Pipe Sandstone at Victoria Falls, Rhodesia," by F. Dixey; (3) "The Geomorphology of Northern Rhodesia," by F. Dixey; (4) "The Bushveld Granites in the Zaaiplaats Tin Mining Area," by C. A. Strauss and F. C. Truter; (5) "The Basal Beds of the Transvaal System at Olifants River Poort, North East Transvaal," by J. W. Brandt and H. D. le Roex; (6) "Columnar, Conical, and Other Growths in the Dolomites of the Otavi System, S.W.A.," by C. M. Schwellnus and H. D. le Roex; (7) "Further Examples of Syntexis by Karroo Dolerite," by E. D. Mountain; (8) "The Origin of the Iron and Manganese Deposits in the Postmasburg and Thabazimbi Areas," by J. E. de Villiers; (9) "Stylolitic Solution in Witwatersrand Quartzites," Robert B. Young; (10) "Further Examples of Subterranean Subsidence of the Marievale Type and Syntexis Associated with Them," by J. Ellis; (11) "The Structure, Ore Genesis, and Mineral Sequence of the Cassiterite Deposits in the Zaaiplaats Tin Mine, Potgietersrust District, Transvaal," by P. G. Söhnge; (12) "The Vredefort Structure as Revealed by a Gravimetric Survey," B. D. Maree; (13) "Stratigraphic Features and Tectonics of Portions of Bechuanaland and Griqualand West," by D. G. L. Visser; (14) "Geomorphology of the Natal Drakensberg," by L. C. King; and (15) "A Comparison of the Gravimeter and Torsion Balance Methods of Geophysical Prospecting," by J. F. Enslin.

"On the Younger Pre-Cambrian Granite Plutons of the Cape Province." By D. L. SCHOLTZ. (Anniversary Address by the President.) (*Geol. Soc. South Africa Proc.*, vol. 49.) Johannesburg, 1947. Pp. xxxv–lxxxii.

Dr. Scholtz's presentation covers the following topics: texture and mode of weathering of the granites, age of emplacement of the plutons, metamorphism of the Malmesbury sediments, inclusions in the granites, mineralogical composition of the granites and quartz porphyries, the structure of the plutons, differentiation and contamination of the granites, radium content of the granites, and the ore deposits. Use is made of Niggli's Q.L.M. diagram in the interpretation of the differentiation of the plutons, the conclusion being that gravitational differentiation, in conjunction with contamination, accounts for the increased basicity with depth. Evidence for a considerable amount of stoping is presented, and Wahl's ideas on thermal diffusion-convection differentiation are considered in accounting for the silica-alkali-alumina-rich border facies of the masses.

Fifty-two new chemical analyses are presented: forty-one of the pluton materials proper, five of contaminated and xenolithic material, and six of the various country rocks. Tables of recalculated chemical data, geologic and structural maps, photomicrographs and outcrop photographs, and variation diagrams are all presented.

In addition to the anniversary address, the 1946 volume of the *Transactions and Proceedings of the Geological Society of South Africa* contains the following papers: (1) "The Discovery and Prospecting of a Potential Gold Field near Odendaalsrust in the Orange Free State, Union of South Africa," by A. Frost, R. C. McIntyre, E. B. Papenfus, and O. Weiss; (2) "Notes on the Microscopic Features of the Magnetic Iron Ores of the Bushveld Complex," by C. A. Strauss; (3) "Corundum 'Indicator' Basic Rock and Associated Pegmatites in the Northern Transvaal," by J. W. Brandt; (4) "The Geology of a Portion of the Rooiberg Tinfields," by L. G. Boardman; (5) "Qualitative Spectrochemical Analysis of Minerals and Rocks," by L. H. Ahrens and W. R. Liebenberg; (6) "Petrofabric Analysis of the Bushveld Gabbro from Bon Accord," by J. J. van den Berg; (7) "Note on a Mechanical Method for the Quantitative Estimation of Graphite," by R. J. Adie; (8) "The Simple Dykes and Sills of the Far East Rand," by J. Ellis; (9) "The Development of the Vaal River and Its Deposits," by H. B. S. Cooke; (10) "Résumé of the Geology of the Richtersveld and the Eastern Sperrgebiet," by P. G. Söhnge and John de Villiers; and (11) "Die Korrelasie van Sekere Voor-Transvaal-Gesteentes in die Distrik Schweizer Reneke," by O. R. van Eeden.

JULIAN R. GOLDSMITH

GLACIOLOGICAL WORK OF THE BRITISH JUNGFRAUJOCH RESEARCH PARTY

"A Crystallographic Investigation of Glacier Structure and the Mechanism of Glacier Flow." By M. F. PERUTZ and GERALD SELIGMAN. (*Royal Soc. London Proc.* ser. A, no. 950, vol. 172.) Pp. 335–360, 1939.

"The Temperature, Melt Water Movement and Density Increase in the Névé of an Alpine Glacier." By T. P. HUGHES and GERALD SELIGMAN. (*Royal Astron. Soc. Monthly Notice, Geophys. Supplement*, vol. 4.) Pp. 616–647, 1939.

"The Structure of a Temperate Glacier." By GERALD SELIGMAN. (*Geog. Jour.*, vol. 97.) Pp. 295–317, 1941.

"Mechanism of Glacier Flow." By M. F. PERUTZ. (*Phys. Soc. Proc.*, vol. 52.) Pp. 132–135, 1940.

"Growth of Glacier Crystals." By GERALD SELIGMAN. (*Nature*, vol. 161, no. 4091.) P. 485, March 27, 1948.

During the past decade and in spite of wartime interruption, British glaciological work has advanced rapidly, as attested by the birth of the *Journal of Glaciology* in January, 1947. Demorest's tragic death in 1942 was a severe setback to American work on the physics of glaciers, but current interest in ice and snow is mounting steadily in the United States and Canada. The value of a review dealing with data published nearly ten years ago and partly digested in Matthes' (1942*a*) excellent treatment on glaciers and in his earlier review of one of the papers (Matthes, 1942*b*) may not be apparent. However, it is felt that the full significance of the Jungfrau research work has been obscured by World War II and that a more complete recapitulation of the major findings may serve to kindle enthusiasm for further work in this promising field on this side of the Atlantic.

British contributions in glaciology during the last decade are largely the product of a research group organized and directed by Mr. Gerald Seligman which takes its name from the Jungfraujoch, an ice- and snow-covered saddle, at 3,460 meters altitude, in the névé region of the Great Aletsch Glacier of Switzerland. This site was selected for a research station because, in addition to its obvious natural fitness, easy access was afforded by the Jungfrau railway leading to a hotel on a near-by rocky peak (the Sphinx), which provided comfortable accommodations for the research party. A meteorological station atop the Sphinx constituted a further asset, and the sum total of these advantages unquestionably allowed this group to operate more effectively than could investigators working in a remote region where equipment is necessarily limited and a major share of the party's time and energy is devoted to the day-to-day problem of staying alive. A cavern excavated in the ice apron of the Sphinx and lighted by power brought from the railway served as a cold laboratory in which to study the physical properties and crystallography of ice and firn. Shafts and bore holes were sunk in near-by firn fields, principally the Mönchfirn at 3,460 meters altitude, and additional investigations were conducted in crevasses of the névé and in ice grottos of the glacier tongue. In none of these did the depth of penetration exceed 30 meters. Mean annual air temperature on the Mönchfirn is $-7°$ C.

For approximately 5 months between April and September, 1938, field investigations were carried forward on the following: (1) transformation of firn into glacier ice, (2) measurement of temperature distribution and variation in the névé area, (3) behavior and function of meltwater in firn, (4) measurements of firn density, (5) evaluation of the function of meltwater and compaction-settling in increasing firn density, (6) crystallographic studies of firn and glacier ice by petrographic methods, (7) mechanics of firn and glacier flow, (8) origin of ice bands in névé and of blue bands in glaciers.

Crystallographic studies by M. F. Perutz showed that most ice crystals in the uppermost layers of the firn were oriented with their *c*-axes nearly vertical, that is, parallel to the temperature gradient. This orientation developed from a dispersed arrangement in new-fallen snow by melting, recrystallization, and sublimation during the transformation from snow to firn because crystals oriented with their *c*-axes parallel to the temperature gradient grew at the expense of others not so oriented. It was also established that at a depth of 14 meters in the firn the vertical orientation had noticeably deteriorated and at a depth of 23 meters it was almost completely destroyed. At still greater depths the dispersed orientation was augmented. However, crystals in the ice bands of the firn maintained a predominant vertical orientation throughout the

depth of observation, and this was attributed to the immobility of the tightly grown crystals composing the bands. Loss of orientation in the loose firn was attributed to shifting of individual crystals or groups of crystals as compaction by settling occurred. This was confirmed by rectangular networks of pins embedded in the walls of shafts. These pins showed movements within the firn to be erratic in direction and amount and to be confined to individual crystals and groups of crystals. No evidence of movement along discrete slip planes was found, and lack of noticeable increase in grain size was thought to rule out recrystallization as a possible explanation for the development of a dispersed crystallographic orientation.

From thermocouples buried in bore holes and in the walls and floor of a shaft it was ascertained that firn temperatures are practically always zero below 15–20 meters, and the conclusion was reached that below these depths the entire glacier is at the pressure-melting temperature. Amelioration of the winter cold wave in the firn and changes of temperature therein were so rapid that they could not have been due wholly to conduction and were, therefore, attributed to percolation of meltwater. Thus, warming of the firn depended primarily on the supply of meltwater and the permeability of the firn as controlled by grain size, packing, density, impermeable crusts, and, above all, horizontal ice bands. By the end of summer, pressure-melting temperatures prevailed throughout the entire névé of the Aletsch Glacier.

Meltwater circulation in the firn was studied in some detail. Pans buried in firn near the surface collected more water than did those at greater depth unless local irregularities, such as ice bands, interfered. As expected, the amount of radiation falling on the surface proved to be the major factor controlling the supply of meltwater, but a fresh fall of snow also produced an increase, owing to the relatively rapid melting of the delicate snow crystals. During warm weather vertically descending meltwater at a depth of 1 meter amounted to 0.1–0.3 cc/cm²/hr. Flow of meltwater along firn layers dipping 8.5° was measured at 8.3 cm. per hour; but its amount, 0.005 cc/cm²/hr, was so small that the accuracy of the volume determination was questioned.

Density in the Mönchfirn at 3,455 meters altitude increased progressively from 0.3 in old snow at the surface to 0.6 in firn at 7 meters depth. Subsequent changes were much slower, the density reaching only 0.68 at 30 meters depth. In a crevasse at 3,300 meters altitude, the firn at 8 meters depth and age two years had a density of 0.63; at 14 meters and three years age the density was 0.72; and at 23 meters depth and 6 years age the density was 0.80. The density of firn is increased by freezing of meltwater and by compaction-settling accompanied by some sublimation and surface migration of water molecules. In general, early-summer freezing of descending meltwater which destroys the winter cold wave was considered more important than early-winter freezing of meltwater remaining in the firn from the preceding summer. The number of times a layer of firn is exposed to seasonal temperature changes before it becomes protected from such changes by deep burial is clearly significant. This is controlled by the annual rate of snow accumulation and depth of penetration of the winter cold wave. If accumulation is small and depth of penetration large, a firn layer will undergo many more seasons of freeze and thaw than if the reverse is true, and the increase of its density by freezing of meltwater will be correspondingly great. On the Mönchfirn a layer of firn undergoes only four or five seasons of freeze and thaw at the most.

Measurements at the beginning and end of the summer at a depth of 1.2 meters showed a density increase from 0.35 to 0.53. This was much greater than the increase accounted for solely by freezing of meltwater, and the difference was attributed to slow settling and compaction. The settling rate of new snow at a depth of 20 cm. was as much as 0.8 cm/cm/day when temperatures were above zero and about 0.13 cm/cm/day when temperatures were below zero. These rates of settling would produce an increase in density of new snow from 0.15 to 0.25 in 12 hours at temperatures above zero or in 3 days at temperatures below zero. The mean rate of settling of firn to a depth of 2.5 meters proved to be $1.9–2.25 \times 10^{-3}$ cm/cm/day. This is sufficient to produce an increase from 0.4 to 0.52 in 100 days. These observations provide the basis for concluding that settling and packing by shifting of crystals is the major cause of density increase in snow and firn; and the fact that no great increase in crystal size is noted, except in the initial stages, is cited as substantiating evidence that freezing of meltwater is a subordinate factor. Furthermore, calculations show that freezing of meltwater can account for only a small part of the total increase in density.

Two types of bands were recognized in the firn—the brownish annual bands and thin strata of clear ice of modest areal extent, one being traced over an area 150 × 160 meters. The annual bands are discolored by fine rock dust, spread over the glacier by wind in late summer, when bare rock is exposed on near-by mountain sides. In the course of excavating and maintaining shafts, it was discovered that the clear ice bands are formed within firn along relatively impermeable, high-capillarity crusts or layers which become saturated with downward-percolating meltwater and are frozen to solid ice by the next cold wave from the surface.

Firn is granular, compacted snow that is distinguished from glacier ice by intercommunicating air channels, which render it permeable to water. In glacier ice the intercommunicating air channels have been broken up and sealed off, so that the mass is relatively impervious. This difference is clearly demonstrated by Perutz's thin sections, which also show that air spaces in glacier ice are fewer, larger, and chiefly intracrystalline. The transition from firn to ice takes place with remarkable uniformity at a density between 0.82 and 0.84 and is not accompanied by a marked rise in density or crystal size.

Ice crystals in firn are rarely larger than 2 mm., but in the glacier tongue they are 1–10 cm. and occasionally more. The dispersed crystallographic arrangement in firn is also replaced by a preferred orientation in the glacier with the basal glide plane of many crystals lying parallel to the direction of flow. An increase in density to about 0.92 also occurs and is attributed largely to growth of crystals with elimination of air bubbles by plastic deformation during flowage. Crystal growth is attributed to the fact that crystals oriented with the basal glide plane parallel to the direction of shear have smaller internal stress and less free energy than do crystals of unfavorable orientation. The former therefore grow at the expense of the latter by transfer of molecules across crystal boundaries. Crystallographic studies showed no evidence of twinning or formation of new crystal nuclei during flow; and no trace of saline films was recorded, although the methods used were capable of detecting concentrations as low as 0.004 per cent NaCl and 0.008 per cent NH_4NO_3.

Lines of screws embedded in the ice wall of a grotto in the glacier tongue recorded actual slip along a blue ice band. Intracrystal air bubbles in ice adjoining the blue band were flattened, and the blue band itself consisted principally of large crystals from which air bubbles had been eliminated and concentrated in a thin layer just above the blue band. Most of these crystals were oriented with the basal glide plane parallel to the plane of the band. Blue bands were observed cutting sedimentary ice bands inherited from the névé, and the general conclusion was reached that they are of tectonic origin. From various crystallographic observations it was concluded that flowage in glacier ice is due to the plastic deformation and growth of individual crystals and to slip along fractures where the stress is sufficient.

Following wartime interruption, work is being continued by this group, with attention focused at present on the mechanism of crystal growth in glacier ice. It is felt that eventual solution of this problem will result in a fundamental contribution to the understanding of glacier flowage. Initial investigations show that crystals in flowing ice are not nearly so large as has been stated heretofore. Further publications from this ambitious glaciological program are awaited with anticipation.

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ROBERT P. SHARP